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Errata

MAIK Nauka regrets a printing error that occured in Fig. 2 of the paper by N.N. Kuz'mina *et al.* [*Stratigraphy and Geological Correlation,* vol. 5, no. 5, 1997, pp. 478–489]. In the lower part of prospect trench 110, at the depth of 5 m, Bed XI, the absolute age date should be 128 000 ± 32 000 instead of 109 000 ± 27 000. Our apologies to the authors of the article and the readers of the journal.

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Cretaceous Evolution of Continental Margins as Related to Global Events

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Abstract—The character and intensity of Cretaceous events in continental margins are genetically linked with the rate of oceanic crust accretion and intensity of endogenic (plume) activity of the Earth. This is most evidently proved by extraordinary events that occurred in continents in the 120–80 Ma interval as dependent on the accretion dynamics of the oceanic crust. The peak accretion was manifested in the Middle Cretaceous tectogenesis of unusual intensity, and a certain attenuation of this process caused more rapid subduction and higher volcanic activity in continental margins. The intense accretion of Cretaceous oceanic crust resulted in the growth of continental crust due to the increased area of continental margins and their greater thickness. The interval of 120–80 Ma is also a period of significant paleoclimatic and biological changes. The plate tectonic activity generally depends on the intensity of the lower mantle upwelling. Stages of intensive mantle plumes correlate with episodes of lithospheric plate reorganization and higher orogenic and magmatic activity, while their lower activity slows down the lithospheric processes. Against the background of continuous plume activity during the Phanerozoic, the interval of 120–80 Ma can be identified as a period of superplumes responsible for extraordinary consequences. Two Cretaceous superplumes, i.e., the Pacific and Atlantic–African, were of different intensity, dynamics, and duration, and this ultimately defined the difference between their boundaries with continental margins, convergent in the Pacific and passive in the Atlantic Ocean. The synchronous activity of plumes resulted in maximum intensity and a global character of events at the lithospheric level (120–80 Ma), whereas the discordant phases of superplumes were manifested by tectonic and magmatic processes of different types in the oceans and continental margins (80–50 Ma).

Key words: *Cretaceous, obduction, subduction magmatism of active continental margins, passive margins, oceanic crust accretion, mantle plumes*

INTRODUCTION

Correlation of geologic events on a global scale is important for establishing their possible synchronism and degree of genetic interdependence. From this point of view, the Cretaceous period in the Earth's evolution is of particular interest, because the Phanerozoic history includes several unique phenomena. Correlation of these phenomena (usually some of them) has attracted the attention of many scientists, who frequently considered them from opposite standpoints concerning the correlation procedure and origin of events (Zonenshain and Kuz'min, 1983, 1992; Lomize, 1986; Bogdanov, 1988; Ronov *et al.*, 1989; Pushcharovskii *et al.*, 1990; Basov and Vishnevskaya, 1991; Larson, 1991a, 1991b; Pushcharovskii and Melankholina, 1992; Khain and Balukhovskii, 1993; Khain and Seslavskii, 1994; Fukao *et al.*, 1994; Kawakami *et al.*, 1994; Maruyama, 1994; Stein and Hofmann, 1994; Zharkov *et al.*, 1995; Milanovskii, 1995, 1996; Khain, 1995; Chumakov *et al.*, 1995; Cambray and Cadet, 1995; Filatova, 1996; and others).

The purpose of this paper was to establish how the stages in the Cretaceous history of continental margins correlate with events in the oceans and with endogenic activity of the Earth as a whole. Vaughan (1995) already demonstrated the effect of endogenic processes of the Earth on Cretaceous orogenic movements in the Pacific continental margins. This effect, however, seems to be of a wider significance; it is confined not only to the Pacific, but also involves a wider spectrum of Cretaceous events in continents.

SPECIFIC FEATURES OF CRETACEOUS HISTORY OF THE ACTIVE CONTINENTAL MARGINS

During Cretaceous time, the active continental margins rimmed the Pacific and were typical of the northern periphery of the Tethys. Apart from certain regional differences in the evolution of continental margins around the Pacific and, in particular, facing two different oceans, the global system of the Cretaceous margins generally shows a common tendency of development (the sequence and intensity of orogenic, metamorphic, and magmatic processes) and coincidence in time of geological rearrangement phases and other most important events. The problem is discussed below by using examples of individual segments of the Creta-

ceous convergent boundaries between oceanic and continental plates.

There were two supercontinents in the Jurassic–Neocomian time: Laurasia and Gondwana. At this stage, a series of island arcs with fore-arc and back-arc basins was formed in the northeastern Pacific near the coasts of western Laurasia. The segments of these island arcs, such as the Yukon–Koyukuk, Toghiak, and Grawina–Natsotin structures, are now recognized as terranes in Alaska and the Columbian Cordilleras (Miller and Hudson, 1991; Samson *et al.*, 1991; Wilson *et al.*, 1991; Currie and Parrish, 1993; Monger, 1993). These terranes include intrusive and island-arc volcanic rocks 137–125 Ma old as determined by the K–Ar method (Miller and Hudson, 1991). Similar island-arc segments, now considered as terranes, are also observed further south along the Oregon–California margin of the United States (Bogdanov and Dobretsov, 1987; McLanghlin *et al.*, 1988; and others). The Rogue terrane, composed of andesites and their tuffs, is 157–150 Ma old, and the K–Ar age of island-arc intrusive rocks of gabbro–granodiorite differentiated series from the Chetco and Rum Creek terranes ranges from 160 to 153 Ma, while the U–Pb dating of zircons puts them in the interval of 157–155 Ma. This Jurassic–Neocomian island-arc system also included the easterly located terrane of the Smartville magmatic complex. The upper age limit of island-arc rocks in the Klamath Mts. is about 148–132 Ma (Frazier and Schwimmer, 1987). The Nevadian (pre-Tithonian) orogeny, which was active 150–145 Ma ago (Harper and Wright, 1984), caused collision between some of these Jurassic island arcs and coupled fragments of Paleozoic–Early Mesozoic island arcs (Peninsular, Wrangellia, and Alexander) into a single Talkeetna superterrane. Their coalescence occurred considerably further south of the modern position of this superterrane (Jones *et al.*, 1986; Wallace *et al.*, 1989). The Nevadian orogeny, however, was of a limited intensity, and the Jurassic island arcs continued their development in the Early Cretaceous as well (Lanphere *et al.*, 1968; 1978).

The Middle Cretaceous orogeny (120–100 Ma) had an enormous impact on the western part of North America. It is called the Columbian orogeny in the northwestern area of the continent and the Sevier orogeny in Oregon and California (Csejtev *et al.*, 1982; Jones *et al.* 1982; Jones *et al.*, 1986; Underschultz and Erdmer, 1991). It was initiated by a more rapid reciprocal convergence of the Farallon and Laurasia plates. This convergence caused an intensive collision of the Jurassic–Neocomian island-arc system with the western margin of Laurasia, and the island arcs thrust over the continent in combination with the oceanic and marginal sea ophiolites (Shervais and Kimbrough, 1985). Consequently, a thrust-nappe orogenic belt of the terrane type was formed and accreted to the continental margin. At that stage, the associated subduction and magmatism attenuated. The time of Middle Cretaceous orogeny manifestation is proved by stratigraphic and age data for metamorphic rocks. In Alaska, fragments of the Jurassic–Neocomian volcanites are found in Albian conglomerates of the Brooks Range overlying the Middle Cretaceous thrust-nappe structures as an intermediate neoautochthone. In northwestern Canada, significant unconformities are observed in the middle Aptian sequence and also between the Albian and Cenomanian beds (Dixon, 1993). The Middle Cretaceous overthrusting of rock complexes onto the continent was coeval with the greenschist–amphibolite metamorphism that affected the tectonic nappes, whose age is estimated to be 112 ± 11 Ma in the Alaska–Columbia–Cordillera margin and about 120 Ma in the Oregon–California margin (Lanphere *et al.*,, 1978; Armstrong *et al.*, 1986; Bogdanov and Dobretsov, 1987).

The Middle Cretaceous collision–obduction processes were also manifested in Central America (Aubouin and Blanchet, 1981; Chekhovich, 1993; Stein *et al.*, 1994; Tardy *et al.*, 1994; Vaughan, 1995).

Owing to the Farallon plate subduction beneath the accretionary western margin of Laurasia, a marginal volcanic belt was formed in this area in the late Albian–Cenomanian, although its magmatic activity was represented mainly by intrusions. This event was associated with the compression regime that appeared as a result of oblique subduction. The Albian–Late Cretaceous plutonic belt ceased to evolve about 86 Ma ago, when the convergent situation changed to transform relations between the oceanic and continental plates (Jones *et al.*, 1986; Coney, 1987).

The Laramide tectogenesis was clearly manifested during the interval of 80–50 Ma in the western part of North America (Jones *et al.*, 1986; Coney, 1987; Wallace *et al.*, 1989). It caused an uplift of the continental margin, complicated structural patterns of the Middle Cretaceous orogenic belt, was responsible for overthrusting of tectonic nappes further to the east, and resulted in the appearance of a chain of depressions with coal-bearing molasse in frontal areas of the nappes.

In the Late Jurassic–Neocomian, a complicated system of marginal–continental and island-arc volcanic belts was formed along the western margin of Gondwana and in the adjacent southeastern part of the Pacific. The northern segment of this system (Colombia, Ecuador, Peru) included an ensimatic island arc (Walrabe-Adams, 1990; Megard, 1987), which further south merged into the Chile–Argentina marginal volcanic belt with thick sequences of calc–alkaline rocks (Lomize, 1983) including diorites and adamellites 130–120 Ma old (Coira *et al.*, 1982). In the back-arc trough, tholeiitic and alkaline magmatic rocks 128 ± 5 Ma old were formed (Grier *et al.*, 1991). The same structures stretched to the south into the Patagonian segment and Antarctica region. The back-arc Rocas Verdes sea basin in Patagonia was opened because of spreading activity, and a segment of the ensialic volcanic belt was detached from the continent (Dalziel *et al.*, 1974; Suarez, 1979; Stern *et al.*, 1991; Wilson, 1991).

Therefore, the formation of Late Jurassic–Early Cretaceous volcanic belts in western Gondwana and western Laurasia occurred simultaneously with crustal extension in continental margins, as is evident from the appearance of back-arc and marginal sea basins with oceanic crust.

The Aptian–Albian (Peruvian) tectogenesis stimulated thrusting and obduction of oceanic, island-arc, and back-arc complexes onto the western margin of Gondwana. The formation of this orogenic belt was associated with the appearance of high-pressure metamorphic zones 115–107 Ma old (McCourt *et al.*, 1984). The Middle Cretaceous orogeny resulted in attenuation of suprasubduction magmatism, in warping and uplift of the western margin of South America, and in formation of the Andean proto-Cordillera.

In the late Albian after the end of the orogenic phase, a new stage of oceanic plate subduction under the accretionary continental margin began and initiated formation of the corresponding volcanic belt in the continental margin. In the peri-oceanic zone of the Pacific, an ensimatic island arc synchronously developed in the frontal part of the northwestern segment of this belt (Wallrabe-Adams, 1990; Aguirre, 1992). The orogenic movements at the Cretaceous–Paleogene boundary (analogues of the Laramide tectogenesis) thrust the island arc over the northwestern margin of South America, where a nappe belt appeared with local zones of the amphibolite–greenschist metamorphism 75–57 Ma old (McCourt *et al.*, 1984). Manifestations of the Laramide orogeny are observed in the entire area of western South America. Here, as in North America, they were less intensive than during the Middle Cretaceous tectogenesis.

In the Late Jurassic–Early Cretaceous, a complicated system of island arcs and microcontinents was located in the northwestern part of the Pacific Ocean east of the Siberian continent (Parfenov, 1984; Filatova, 1988; Zonenshain *et al.*, 1990; Sokolov, 1992). It developed until the Aptian despite the short Late Jurassic (Kolymian) orogeny, which caused collision of continental blocks and island arcs with the Siberian continent and formation of high-pressure metamorphic zones with anatectic granitoid bodies about 140 Ma old (Parfenov *et al.*, 1993). Further to the south, at the China–Korean edge of Laurasia, the island arcs adjacent to the continental margin were transformed into the Late Jurassic–Neocomian volcanic belt 134–11 Ma old (Wang, 1985; Wu and Pei, 1988; Filatova, 1991).

About 125–120 Ma ago (end of the Barremian–beginning of the Aptian), the eastern margin of Laurasia experienced intensive compression, owing to which the subduction zones together with associated continental margins and peri-oceanic island-arc belts ceased their activity. The Middle Jurassic orogeny resulted in collision of island arcs and microcontinents with the Siberian continent and produced anatectic granitoids 120–105 Ma old in the collision zones (Parfenov *et al.*, 1993).

In the course of the Middle Cretaceous orogeny, terranes of different origin, such as island arcs, marginal sea basins, and oceanic formations, migrated from the Pacific Ocean and collided with the northeastern margin of Asia (Parfenov *et al.*, 1993; Filatova and Vishnevskaya, 1997). As a consequence, the Western Kamchatka–Koryak accretionary orogenic belt was formed and enlarged the area of the continental margin of Asia. The southern continuation of the Middle Cretaceous orogenic thrust–nappe belt is traced in the Khabarovsk and Amur rock complexes of western Sikhote Alin' and also in the inner nappe zone of southwestern Japan (Kojima, 1989; Natal'in and Borukaev, 1991).

In the late Albian, the collision–accretion processes at the edge of the Asiatic continent changed to rapid subduction of the Kula plate, and the extended Chukotka–Cathasian volcanic belt consisting of many segments appeared in the eastern margin of Asia. The northernmost Albian–Cenomanian Okhotsk–Chukotka segment unconformably overlapped the edge of the ancient continent and adjoining Middle Cretaceous thrust–nappe structures of the West Kamchatka–Koryak tectonic zone (Filatova, 1995). To the south, the continental margin was marked by the Khingan–Okhotsk segment (Natal'in and Borukaev, 1991), which further south merged with the Cathasian segment of the Albian–Late Cretaceous volcanic belt.

The Laramide tectogenesis was manifested also along the western continental margins of the Pacific Ocean, but not as intensively as in the western part of North America. Areas of orogenic contraction alternated here with areas of extension, which hosted the intraplate basaltoid magmatism (Filatova, 1988; 1995).

In the Tithonian–Hauterivian time (155–130 Ma), as a result of oceanic plate subduction under the Tethyan margin of Laurasia, the Pontian–Caucasian ensialic island arc was formed (Zonenshain *et al.*, 1987). Behind it, the proto-Black Sea and South Caspian marginal basins appeared. The Late Jurassic orogeny (an analogue of the Nevadian phase) in the Mediterranean region is marked by late Kimmeridgian folding in a considerable segment of the South Eurasian margin extending from Dobrudjia to the Caucasus inclusive. According to reconstructions for the end of the Jurassic and beginning of the Cretaceous, a system of left-hand displacements in the Tethys area was active, and the most intensive movements occurred in the northern part of the African–Arabian plate. The Barremian–Albian interval (130–110 Ma) was a time of collision events (the Austrian tectogenesis) at the northern borderland of the Tethys (Knipper, 1985; Zonenshain *et al.*, 1987; Kaz'min, 1989; and others). A partial closing of the Ligurian basin and obduction of the northern part of the Apulian block together with ophiolites onto the Eurasian margin initiated formation of the Austrian–Alpine nappes. At the same time, in the northern part of

the African–Arabian plate, an oceanic basin appeared as a result of transformation of rift structures. During this stage, several microcontinents were accreted to the Eurasian margin (central Afghanistan, the southern Pamirs, and southern Tibet) along the corresponding sutures (Varash, Rushan–Pshart, Bangkok; Tapponier *et al.*, 1981; Otsuki, 1985; Zonenshain *et al.*, 1987). The Middle Cretaceous collision events caused the intense folding and thrusting of rock sequences in the southern margin of Eurasia, where structures display the distinct northern vergence; this stage also resulted in a temporary cessation of oceanic crust subduction under this margin and low activity of the Jurassic–Neocomian volcanic belt (Kaz'min *et al.*, 1987). Some volcanoes, e.g., those indigenous to the Trans-Caucasus region, remained active.

The Albian–Late Cretaceous (100–80 Ma) is recognized as a stage of intense tectonic rearrangement within the Tethys area (Knipper and Sharas'kin, 1995). In its northern part, a new subduction zone dipping under the accretionary margin of Eurasia appeared together with a volcanic arc above it; the latter was in fact the marginal volcanic belt located on the continental edge covered by sea. Behind this zone, a system of deep-water basins was still in existence. The volcanic belt included the Lesser Caucasus and extended further east across central Afghanistan, the Pamirs, and southern Tibet up to Indochina (Otsuki, 1985; Kaz'min *et al.*, 1987; Peltzer and Tapponier, 1988). In addition, suprasubduction magmatism in the Late Cretaceous was activated owing to formation of the East Carpathian and Balkan basins. It is important to note that the Albian–early Campanian calc–alkaline volcanism in the South European active margin represented an independent event isolated from the Jurassic–Neocomian and Paleogene magmatism (Kaz'min *et al.*, 1987). In the Albian–Campanian time, the volcanic activity in continental margins episodically attenuated because of collision of the continental margin with segments of mid-ocean ridges that approached it from the south and with microcontinents broken off Gondwana (Otsuki, 1985; Zonenshain *et al.*, 1987; Kaz'min, 1989).

At that time, sinistral displacements of continental blocks were still active in the northern part of the African–Arabian plate and southeastern Eurasia. The opening of the southern branch of the Tethys (Mesotethys) that began in the Early Cretaceous also continued in the Late Cretaceous. The Cyprean ensimatic island arc of the Mesotethys appeared in the early Cenomanian (Knipper, 1985; Kaz'min, 1989).

From the beginning of the late Campanian, Africa started its movement toward Eurasia (Zonenshain, 1987; Knipper and Sharas'kin, 1995). The related Laramide tectogenesis resulted in cessation of spreading in the southern branch of the Tethys, in collision of the Apulian block with the African–Arabian plate, and in collision of the latter with the Cyprean arc, whose activity ceased in the Maestrichtian (Zonenshain *et al.*, 1987). Tectonic nappes of this stage were formed in the Alpine region and Dinarides. In the southern margin of Eurasia, including the Carpathians, the calc–alkaline magmatism almost ceased, and, in the late Campanian–Maestrichtian, subalcaline basaltoids were locally accumulated (Kaz'min *et al.*, 1987). The latter closely resemble the Maestrichtian–Paleogene basaltoids of the intraplate type erupted in the western framing structures of the Pacific. At the Cretaceous–Paleogene boundary, the peculiar feature of the Laramide tectogenesis in the Mediterranean was a considerable growth of back-arc basins near the Eurasian margin in the Crimea–Caucasus region; their ensemble is considered as the Paratethys. At this stage, areas of compression and orogeny in the northern margin of the Tethys alternated with areas of extension. It is worth noting that basaltoid magmatism of the intraplate type, the indicator of crustal extension, was also active at the Cretaceous–Paleogene boundary in the Indian region, where the Deccan traps were formed 65–59 Ma ago (Bhale, 1983).

CORRELATION OF CRETACEOUS STAGES IN THE EVOLUTION OF ACTIVE CONTINENTAL MARGINS WITH OTHER GLOBAL GEOLOGIC EVENTS

Among the unique Cretaceous events in active continental margins, the most important were the formation stages of global orogenic belts of the terrane type 120–100 Ma ago and the appearance of the world system of volcanic belts in suprasubduction zones along continental margins 100–80 Ma ago. These events were coeval with other global phenomena, the most important of which were the reorganization of lithospheric plates, paleoclimatic and biologic changes, and specific features of the pulsating endogenic (plume) activity of the Earth.

Correlation of Events in the Continental Margins and Oceans

In the Cretaceous time, evolution of continental margins highly depended on the Pangea break-up and, primarily, on disintegration of Gondwana. The latter turned out to be divided into South America, Africa, India, Australia, and Antarctica. Intense rifting in continents rapidly led to the formation of extended systems of mid-ocean ridges and accelerated sea-floor spreading (Zonenshain *et al.*, 1984; Scotese *et al.*, 1988; Khain and Balukhovskii, 1993; and others). The interval of 120–110 Ma was a time of opening of the South Atlantic Ocean, and its mid-ocean ridge joined the ridge system of the Central Atlantic 110–85 Ma ago. In the North Atlantic, the spreading zone gradually migrated to the north (Rowley and Lottes, 1988). The spreading activity in the Canadian basin of the Arctic began at the end of the Neocomian and was at its max-

imum 110 Ma ago (Rowley and Lottes, 1988; Grantz *et al.*, 1994).

In the Atlantic Ocean, a stepwise growth of the oceanic crust commenced 118 Ma ago (from 0.24 to 0.54 km^2/year) with a subsequent gradual increase up to the early Eocene, when it reached 0.78 km^2/year (Zonenshain and Kuz'min, 1992). In the Pacific, an anomalously high spreading rate occurred in the interval of 118–80 Ma; in the Barremian, the rate of oceanic crust increment was lower than 15 km^3/year, whereas in the Aptian it reached 20 km^3/year (Larson, 1991a). The extraordinary events on continental margins, therefore, corresponded to the stepwise increase in the production rate of oceanic crust during the Middle–Late Cretaceous (120–80 Ma; see figure), because in the Barremian this parameter was 17 km^3/year and in the Aptian it exceeded 30 km^3/year.

In order to understand the causes of events in the Cretaceous convergent boundaries of lithospheric plates, it is necessary to take into account not only this stepwise impulse, but also its dynamic peculiarities (Larson, 1991a, 1991b). In this respect, the time interval of the anomalously accelerated spreading rate consists of two parts (see figure): the Aptian–early Albian interval of peak values (120–100 Ma) and the subsequent Albian–middle Campanian interval (100–80 Ma) of a gradually decreasing, though generally high, accretion rate of oceanic crust. The late Campanian–Maestrichtian interval was the time of a sudden minimum of crustal accretion in the World Ocean, which was almost identical to that of the Barremian time.

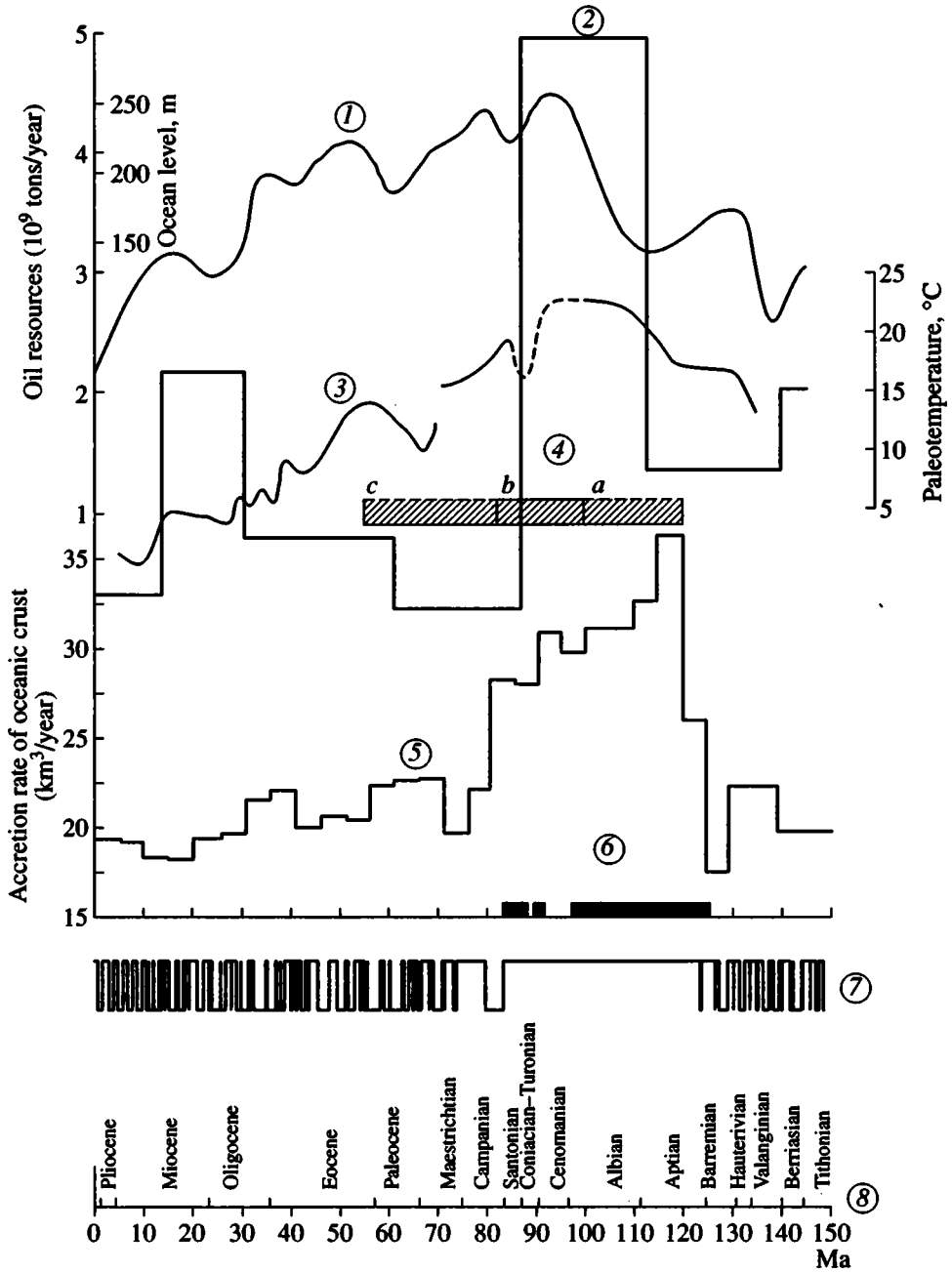
The mentioned features of Cretaceous development of the World Ocean were coeval with events in the active continental margins of the Earth. As stated above, the anomalous peak of oceanic crust accretion in the period of 120–100 Ma is manifested by the extremely intense Middle Cretaceous global orogeny. The tectogenesis of that period developed under conditions of increased crustal compression at the convergent boundaries of continental and oceanic plates. This activity caused accretion–collision processes along the continental margins and their growth owing to accreted terranes of different origin, which collectively composed the orogenic belts of thrust–nappe structure. In the course of thrusting with partial obduction of ophiolite and island-arc complexes onto the continent, zones of high-pressure and low-temperature metamorphism were formed at the base of tectonic sheets. An extremely intense compression terminated the activity of subduction zones in the continent–ocean transition zone and the marginal and island-arc volcanic belts genetically associated with them. This process was also responsible for closing up the marginal and back-arc basins (with the exception of the Black Sea–Caspian depressions), whose rock associations were thrust over the continents.

The processes of extremely powerful Middle Cretaceous tectogenesis, the formation of very complex structures in the Middle Cretaceous orogenic belts, and multiple ophiolite obduction were caused by the intense convergent movements of oceanic and continental plates (Filatova, 1996). These movements were initiated in the interval of 120–100 Ma by the coeval maximum of the spreading rate in all the oceans of the Earth, both in the newly opened oceans and in the Pacific that existed earlier. No other coinciding maxima of oceanic crust accretion are known in the Mesozoic and Cenozoic, and the Middle Cretaceous tectogenesis is outstanding among other Mesozoic–Cenozoic orogenic phases in its intensity and global manifestation.

In the interval of 120–100 Ma, the Middle Cretaceous orogeny was manifested not only by a continental area increase, but also by the growing thickness of continental margins caused by tectonic nappe formation and general warping. These processes, in their turn, resulted in sea regression from marginal zones of the continents and in a sudden reduction of shelf–sea areas. The upheaval of continental margins was accompanied by their intensive erosion and appearance of stratigraphic and structural unconformities.

The second stage (100–80 Ma), when the accelerated accretion of oceanic crust slowed down (see figure), coincided with the termination of Middle Cretaceous tectogenesis and resulted in the appearance of new convergent boundaries and accelerated subduction of oceanic plates under the accreted continental margins. In the course of these events, the Middle Cretaceous stage of attenuation of subduction magmatism was followed by the appearance of a global system of volcanic belts on continental margins during the late Albian, whose activity was unusually high in the Mesozoic and Cenozoic. This activity could serve as an indicator of the high subduction rate during the Albian–Late Cretaceous, because these two processes are tightly interrelated (Lomize, 1986). In the Albian–Senonian time, the eustatic rise of the sea level in the World Ocean (Haq *et al.*, 1987) also resulted in transgression on the continental edges, which is usually thought to be associated with the peak of oceanic crust accretion, volcanic activity of the intraplate type, and sea floor rise in the Pacific. In the Early–Middle Cretaceous, the break-up of Gondwana and Laurasia caused the appearance of equatorial currents that joined the Pacific, Tethys, and Atlantic oceans and were responsible for a more intensive hydrodynamic activity. At the beginning of the Cretaceous, this process in the Pacific was manifested by a maximum of erosion activity not only in the uplifts, but in the deep-water depressions as well (Basov and Vishnevskaya, 1991). Moreover, in the middle of the Cenomanian, the carbonate–siliceous sedimentation changed to deposition of noncarbonate sediments. The accelerated sea-floor spreading in the Pacific was another factor that activated the circulation of water masses.

The late Campanian–Maestrichtian minimum of crustal accretion in the oceans (80–65 Ma) was coeval



Correlation of the most important events in the Earth during the Cretaceous period (after Larson, 1991b, with modifiations): (*1*) curve of eustatic level oscillations in the World Ocean (Haq *et al.*, 1987); (*2*) curve of secular changes in oil production (Irving *et al.*, 1974); (*3*) curve of surface water temperatures in the high latitudes (Arthur *et al.*, 1985; Savin, 1977); (*4*) major Cretaceous events in active continental margins: (a) Cretaceous orogeny, (b) stages of accelerated subduction of oceanic plates under continents and coeval episodes of highly active volcanism in continental margins, (c) the Laramide orogeny; (*5*) curve of secular changes in the global accretion rate of oceanic crust (Larson, 1991b); (*6*) stages of accumulation of black shales (Zharkov *et al.*, 1995; Jenkyns, 1980); (*7*) scale of geomagnetic reversals (Harland *et al.*, 1990); (*8*) geologic time scale (Harland *et al.*, 1990).

with the sudden fall in the convergence rate between oceanic and continental plates and with the slowing down of subduction and correponding reduction of the volume of suprasubduction magmatism up to its cessation in most segments of the Albian–Late Cretaceous system of volcanic belts. The attenuated convergence of oceanic and continental plates created conditions for crustal extension in certain areas of continental margins, which was accompanied by a pulse of intraplate basaltic eruptions at the Cretaceous–Cenozoic boundary.

The indicated processes, therefore, designate a close connection between the events that occurred on continental margins and the sea-floor spreading rate,

the individual features of which were specified above. According to Bott (1993), the accelerated spreading of the oceanic crust activated the ridge–push forces along the periphery of the oceans. The compression regime that appeared at this stage caused a sharp increase of adhesion between the subducting oceanic plates and margins, thus impeding the further development of subduction and suprasubduction magmatism and causing mutual deformations of plate couples, their warping, and rising of the suprasubduction edge of the continent. Vaughan (1995) used this model to account for the Middle Cretaceous orogeny in continental margins.

As shown above, however, the relationship between the accretion rate of the oceanic crust and the Cretaceous evolution of continental margins is more complicated and multiform though dependent on the rate value. The anomalously high value (more than 35 km^3/year; Larson, 1991b) caused the unusually powerful Middle Cretaceous tectogenesis and thrusting of oceanic and island-arc complexes over the continental margins, where the thrust–nappe orogenic belts were formed. The less intense late Albian–early Campanian accretion of oceanic crust (about 30 km^3/year) reduced the compression along the oceanic periphery and compensated the oceanic crust accretion by its accelerated subduction under the continental margins followed by anomalously high volcanic activity. A sudden reduction of oceanic crust accretion at the end of the Late Cretaceous (up to 20 km^3/year) decelerated the subduction and deactivated the global system of volcanic belts in continental margins (with the exception of some of its segments).

It should be emphasized here that the relations between Cretaceous events in continental margins and those in the oceans were much more complicated, because the dynamic development of individual oceans at that time was essentialy different (Zonenshain and Kuz'min, 1992). For example, a sudden reduction of the spreading rate in the Pacific during the Campanian–Maestrichtian coincided with the maximum of oceanic crust accretion in the Atlantic Ocean responsible for the Laramide orogeny. The magnitude of this Atlantic maximum was relatively low, and, consequently, the intensity of the Laramide tectogenesis that developed against the background of the general global reduction of oceanic crust accretion was lesser than that of the Middle Cretaceous orogeny. These conditions in the oceans predetermined the synchronous combination of the Laramide orogenic events with crustal extension and intraplate volcanism in the continental margins.

The events in continental margins were dependent, apart from the spreading rate, on the orientation of movement vectors of plates. When an oceanic plate approached a continent at an acute angle, their boundary was under shear conditions, and magnitudes of orogenic movements and suprasubduction magmatism in the continental edge considerably decreased.

The correlation analysis of events in the oceans and continents shows that the Cretaceous impulse of high oceanic crust accretion (120–80 Ma) correlates with an extraordinary increment of the continental crust. The nature and rates of continental growth in that time interval were different and directly dependent on the crustal accretion rate in the oceans. The maximum increment of continental crust was in the interval of 120–100 Ma, when the expansion of the continental area and greater thickness of continental margins resulted from the accretion–collision processes. Later (100–80 Ma ago), the thickness of continental margins grew owing to the anomalously great volume of suprasubduction volcanic rocks. The thickness of continental crust was likewise increased by the ample volcanic eruptions of the intraplate type in the interval of 125–83 Ma (Stein and Hofmann, 1994).

Correlated Cretaceous History of Continental Margins and Other Global Events

The Cretaceous was a period of several climatic and biological events, which are the object of a separate discussion and which are mentioned here only in relation to the Cretaceous reorganization of lithospheric plates and to magmatic manifestations, including those in the continental margins.

The Cretaceous period was a time of a warm glaciation-free climate with moderately thermophilic plant and animal communities even in high latitudes (Chumakov *et al.*, 1995). The interval of 120–80 Ma corresponded to the peak of global temperatures of surface waters (see figure) as high as 20–22°C in the high latitudes (Savin, 1977; Arthur *et al.*, 1985, 1991). According to the oxygen isotope composition in planktonic foraminifers, the temperature of surface waters gradually grew beginning from the Albian and reached its maximum values in the Turonian–early Campanian time, after which the Campanian–Maestrichtian cooling took place (Naidin *et al.*, 1986; Huber *et al.*, 1995). Consequently, while in the late Albian and late Maestrichtian the difference in the temperature of the surface waters in high and low latitudes was 14°C, in the Coniacian–Santonian it was only 0–4°C.

The Cretaceous warming of the climate is usually attributed to several causes. They could have been the Cretaceous reorganization of continents, the break-up of Pangea and Gondwana, and the appearance of new oceanic areas and a single oceanic equatorial belt (Zonenshain and Kuz'min, 1993; Zharkov *et al.*, 1995; and others). However, since the Cretaceous warming cannot be attributed only to the reorganization of plates (Barron and Washington, 1982), the effect of extremely large amounts of CO_2 in the atmosphere was also suggested. The idea is that this effect was produced by mantle degasing through an extended system of mid-ocean ridges and suprasubduction belts and also through vast mantle reservoirs of intraplate volcanism (Arthur *et al.*, 1991; Calderia and Rampino, 1991;

Maruyama, 1994). It seems apparent that the combined degasing of these magmatic sources was the decisive factor in the ejection of large amounts of CO_2 that formed the greenhouse effect, moreover because the activity of these three sources in the Cretaceous was anomalously high, as compared with other Phanerozoic periods in the history of the Earth. It may be assumed, however, that if the peak of paleotemperatures was in the Cenomanian–early Campanian interval (Huber *et al.*, 1995), then the greater part of carbon dioxide was supplied by the global system of suprasubduction volcanic belts in continental margins, whose exceptionally powerful volcanism occurred in this time interval.

The sudden warming of the climate and abundant vegetation of the Middle Cretaceous were favorable for intensive coal accumulation and formation of black shales, oil, and gas deposits. The large amounts of sulfur and phosphorus extracted from the mantle melts in the course of Cretaceous volcanism of various types were consumed by zoo- and phytoplankton, which are the organic sources of petroleum. The peak of black shale formation falls in the Aptian–Santonian (see figure) and is attributed to several causes: the growth of carbon content, a sudden increase of biomass, and the poor aeration in some basins responsible for an anoxic environment (Jenkyns, 1980). The stagnant conditions and weak circulation of water masses were typical of small and shallow incipient oceans, which were represented by relatively closed basins with restrained water exchange with other basins (Basov and Vishnevskaya, 1991). These environments were settled in the Early Cretaceous in the Central and South Atlantic at the beginning of Pangea break-up. In the Aptian, an intensive accumulation of black shales began in these basins. In the course of further opening of the Atlantic and more active water circulation, the accumulation areas of shales of this type were considerably reduced and almost disappeared by the Campanian time (Zharkov *et al.*, 1995).

The rise of the sea level in the World Ocean began in the late Albian, and, in the Late Cretaceous, this event caused accumulation of another type of black shales in the passive continental margins (Zharkov *et al.*, 1995). This was stimulated here, first, by the rapid burial of large vegetal masses in the shelf seas with greatly expanded areas and, second, by the anoxic environment caused by oxidation of large biogenic masses (Basov and Vishnevskaya, 1991).

As regards petroleum, about 60% of its world resources were generated in the interval of 112–88 Ma (Larson, 1991b). This Albian–Turonian peak of hydrocarbon production was predetermined by the Middle Cretaceous combination of interrelated conditions, such as the high rate of oceanic crust accretion, intensive volcanism, greenhouse effect, high bioproductivity in the oceans and continents, and sudden expansion of shelf-sea areas.

Several other events also had a significant influence on the evolution of marine and land organisms: climatic changes with the temperature maximum in the Middle Cretaceous, reorganization of continental and oceanic plates, appearance of new oceans and global current systems, and the late Albian–Late Cretaceous global transgression. Moreover, the most significant biotic changes occurred at the Early–Late Cretaceous (Naidin *et al.*, 1986; Vishnevskaya and Basov, 1995) and Cretaceous–Paleogene boundaries (Basov and Vishnevskaya, 1991). An important transformation of plant life and flourishing of angiosperms took place in the Middle Cretaceous with concurrent intensive development of insects (Naidin *et al.*, 1986).

Correlation of Events in Continental Margins with Endogenic Activity of the Earth

The Aptian–Campanian high spreading rate in the oceans and coeval events on the active continental margins described above coincided with a quiescent period lacking reversals of the magnetic field in the interval of 125–83 Ma (Larson and Pitman, 1972). This circumstance alone is peculiar in the Earth's history for the last 150 Ma (see figure). In the same time interval, a suddenly increasing volume of intraplate basaltoids (Larson, 1991b) emerged onto the surface of continental fragments of Pangea and onto oceanic plates in the form of oceanic plateaux and guyots. The peak of intraplate volcanism falls within the early Aptian–Albian (120–100 Ma), and its gradual decrease is traced up to 80 Ma. According to geological data, a short impulse of intraplate volcanism also occurred at the Cretaceous–Cenozoic boundary. These events are recognized as indicators of the extraordinary Middle Cretaceous ejection of heat and deep mantle material from the D" layer (the core–mantle boundary) by one or several superplumes (Larson, 1991a; Larson and Olson, 1991; Stein and Hofmann, 1994). The thermal energy of ejections should have changed the temperature gradient at the core–mantle boundary, thus provoking a quicker convection in the outer part of the core, i.e., an accelerated escape of heat from the core compensating the heat loss caused by the mantle upwelling from the seismic D" layer. As soon as this mechanism began, the inversions of the magnetic field of the Earth ceased (Larson and Olson, 1991). By the end of the superplume episode, the temperature gradient returned to its initial value, and the long interval of the normal magnetic field was succeeded by a period of frequent magnetic reversals, which continue up to the present time (see figure).

In a number of models, the upwelling of deep-seated matter and energy along with ceased reversals of the Earth's magnetic field are attributed to the periodic sinking of the cold oceanic lithosphere in subduction zones down to the mantle–core boundary (Honda *et al.*, 1993; Weinstein, 1993; Fukao *et al.*, 1994; Kawakami *et al.*, 1994; Maruyama, 1994; Stein and Hofmann, 1994; Khain, 1995). In the course of subduction, this

matter was accumulated at a depth of about 670 km (the boundary between the upper and lower mantle), where it gradually created a gravity instability, which actually caused periodic sinking of this matter into the D" layer. As was already noted (Pushcharovskii *et al.*, 1990), during certain time intervals, the convective movements in the Earth's interior were of the cascade type.

The presence of not only enriched but also depleted rock varieties in the basaltoid series of the intraplate type can be considered as geochemical indicators of penetration of the ancient subducted lithosphere, which was poor in rare and trace elements, into the D" layer zone (Kerr *et al.*, 1995). The convection mechanism, where the subducted crustal slabs represent the descending branch and mantle plumes represent the ascending branch, suggests recycling of lithospheric and upper mantle matter in combination with the evacuation of undepleted matter from the lower mantle onto the Earth's surface. As a result of this process, even the most depleted basalts of the intraplate type radically differ in geochemical and isotope characteristics from the depleted basalts of the mid-ocean ridges (Zonenshain and Kuz'min, 1983; Stein and Hofmann, 1994; Kerr *et al.*, 1995).

While the idea of sinking of the subduction material to the core–mantle boundary is still problematic, the plume type of magmatic activity is established by a set of geological, petrological, and geophysical data. The above synthesis of data on the Cretaceous history of the Earth allows me to discuss the problem of the relationship between the Cretaceous reorganization of lithospheric plates, including the evolution of continental margins, and the mantle upwelling.

Since, according to geological and geochemical data, the mechanisms of plate tectonics and eruptions of intraplate basalts onto the Earth's surface in the Phanerozoic were active almost synchronously, the concept of mantle plumes rising from the core–mantle boundary becomes rather popular as an addition to the double-layered (lower and upper mantle) convection mechanism. According to this model, the lithospheric plate movements occurred as a result of upper mantle convection, which is, therefore, considered responsible for the opening and closing of the oceans, in other words, for the so-called Wilson cycles (O'Nions *et al.*, 1980; Zonenshain and Kuz'min, 1983, 1992; Stein and Hofmann, 1994). These researchers also emphasized the dominant role of mantle plumes in such significant events in the lithosphere as the Pangea break-up and paroxysms of tectogenesis. This point of view was employed in a model suggesting the succession of episodes of two-layer and one-layer convection systems in the Earth's history (Sheridan, 1987; Stein and Hofmann, 1994). In this model, stages of the two-layer convection responsible for the lithospheric plate movement are isolated from stages of the global single-layer convection caused by the rise of powerful plumes (superplumes) from the core–mantle boundary that involved into the mutual recycling the undepleted lower mantle and depleted upper mantle. At the lithospheric level, the plume stages of single-layer convection cause the break-up of continents, the most intensive orogenesis, crustal accretion of different types, and so on (Stein and Hofmann, 1994).

It is hardly possible, however, to adopt this point of view, because according to it the Wilson cycles (the main phenomena at the lithospheric level) initiated by the two-layer convection are considered separately from the plume stages and, furthermore, are contrasted to them. Since the most important events in plate tectonics coincide with activation of powerful lower-mantle plumes, it is more likely to assume that the plume mechanism actually predetermines the main peculiar features in lithospheric development. This is why the activity of plate tectonics in the history of the Earth primarily most likely depends on the intensity of processes at the core–mantle boundary, i.e., on the activity of plumes rising from this boundary, which, by the way, also affects the activity of the upper mantle convection (Khain, 1995).

It is well known that the plume mechanism of the Earth is always active, as is proved by the nearly continuous age series of Phanerozoic intraplate basalts. The mantle upwelling, however, is unstable in time, as was already indicated (Pushcharovskii *et al.*, 1990). It seems safe to assume that episodes of recurrent endogenic plume activity of the Earth were coeval with all considerable stages of reorganization of lithospheric plates, epochs of intense tectogenesis, and other important events. Some unique superplume episodes stand out against the background of continuous plume activity of different intensity in the Phanerozoic history of the Earth (Larson 1991a, 1991b). The Cretaceous (120–80 Ma) is one of the periods of these superplumes, but it is characterized by a series of catastrophic events unique on the Earth during the last 150 Ma; much earlier, a similar superplume episode occurred during the interval of 323–248 Ma (Larson, 1991b).

The stages of weaker plume activity at the lithospheric level are reflected in a slower sea-floor spreading, decelerated, movement of lithospheric plates and slower subduction, and also in the less intense orogeny and lesser volumes of volcanites of both suprasubduction and intraplate types. Thus, episodes of intensive superplume plate tectonics were followed by periods of slower and sluggish lithospheric processes related to the lower activity of mantle upwelling.

The lithospheric plate tectonics is influenced not only by the intensity of mantle upwelling and the number of active plumes operating simultaneously, but also by the peculiarities of their distribution in space. For the period of 120–80 Ma, only one superplume is usually suggested to be active in the Pacific, covering an area of $6000 \times 10\,000$ km^2 (Larson, 1991a). This exceptionally powerful ejection of deep mantle material

(including intraplate basalts) and heat energy caused the warming and reduction in density of the Pacific lithosphere. It is considered as responsible for a sudden increase in the spreading rate and general uplift of the ocean floor (Schlanger *et al.*, 1981; Bott, 1993). It is this superplume that is usually associated with the tectogenesis in the Pacific continental margins (Vaughan, 1995).

In the Cretaceous, not only the Pacific but also the Atlantic–African superplume was evidently active; it was first identified as a hot field of the Earth by Zonenshain and Kuz'min (1983, 1992). The Atlantic–African superplume of the Cretaceous time was less intense, because the sea-floor spreading within the area of its projection on the Earth's surface was significantly slower than that in the Pacific. Nonetheless, the geological consequences of this superplume were indeed catastrophic and included the break-up of Gondwana, a more rapid development of separate segments of the extended ridge system in the Atlantic Ocean, and their integration into the general global system. An intensive extension in the central and, particularly, southern Atlantic caused the easterly displacement of the African–Arabian continental plate. This movement initiated the origin of a series of sinistral strike-slip faults along the boundary between the Eurasian and African–Arabian plates, and they subsequently resulted in the formation of rift zones in the Tethys province adjacent to this boundary. Later, these zones transformed into spreading zones. Throughout the Cretaceous, the long-term development of this fault system was combined with the direct effect of the Atlantic–African mantle superplume; this caused repeated detachments of continental blocks from the African–Arabian plate and their migration because of spreading in the rear zones towards the Eurasian continent.

Owing to activity of the Atlantic–African superplume, the Cretaceous reorganization of lithospheric plates was supplemented by concentration of the intraplate volcanic fields (Zonenshain and Kuz'min, 1992; Muller, 1993).

A comparison between the Pacific and Atlantic–African superplumes of the Cretaceous time shows that they differ in both the intensity and the time interval of their manifestation. The rate of oceanic crust accretion can serve, of course, as a measure of intensity of manifestations, which in the Pacific was an order higher than that in the Atlantic during the Cretaceous (Zonenshain and Kuz'min, 1992). Consequently, the Pacific superplume was much more intense than the Atlantic–African one, and this caused differences in the Cretaceous history of the continental margins of these two oceans and in the type of continent–ocean boundaries. As shown above, the Cretaceous boundaries between the Pacific oceanic plates and continental margins were convergent. The permanently active compression forces at the periphery of the Pacific were dependent on the spreading rate and realized either as subduction or accretion–collision processes along the convergent boundaries. The relatively low spreading rates in the Atlantic Ocean probably hindered the more or less significant compression in the continent–ocean transition zone, and, consequently, the boundaries of this oceanic basin remained passive. It seems that the slow oceanic crust accretion in the Atlantic was compensated by the gradual moving apart of continental margins without subduction or obduction effects.

The difference in the crustal accretion rates between the oceans (Zonenshain and Kuz'min, 1992) is an indication of the differences in the dynamics and duration of activity of the Pacific and Atlantic–African superplumes. A sudden growth of the oceanic crust volume in the Pacific Ocean during the interval of 118– 83 Ma was followed by the reduction of this parameter almost by half in the Campanian–Maestrichtian, and this shows that the period of existence of the Pacific superplume was relatively brief. The Atlantic–African superplume lasted longer: a tendency of persistent, though not as rapid as in the Pacific, sea-floor spreading can be traced from 118 to 54 Ma with a maximum between the Mesozoic and Cenozoic. Consequently, this superplume, though relatively less intense than the Pacific one, was most active by the end of its existence at the Cretaceous–Cenozoic boundary, when it resulted in the Laramide tectogenesis. The interesting point is that the Campanian–Maestrichtian stage in the Pacific was marked by a burst of intraplate magmatism (Basov and Vishnevskaya, 1991), and this may imply a certain interrelation between both Cretaceous superplumes.

CONCLUSION

(1) The Cretaceous history of the active continental margins of the Earth, which are characterized by unique peculiarities of tectogenesis and magmatism, shows a positive correlation of their development with the crust accretion rate in oceans and with the intensity of endogenic activity of the Earth;

(2) The extraordinary events in continental margins developed against the background of the Pangea breakup and coincided with the interval of steplike growth of the oceanic crust 120–80 Ma ago (Larson, 1991a, 1991b; Zonenshain and Kuz'min, 1992). The character of these events depended on the crustal accretion dynamics within the indicated time interval in the following manner:

(a) The peak of crustal accretion in the oceans (120–100 Ma) was manifested in the most intense compression in their peripheral parts and caused the formation of thrust–nappe orogenic belts together with obduction and metamorphic events in the continental margins. Structures of these belts also incorporate terranes of both continental, oceanic, and island-arc origin. The ridge-push forces at the time of anomalously rapid crustal growth in the oceans obstructed subduction

along their periphery and caused attenuation of the activity of volcanic belts;

(b) A certain decline in the accelerated accretion of oceanic crust during the interval of 100—80 Ma somewhat reduced the compression regime at the periphery of the oceans, and the rapid spreading was compensated by accelerated subduction of the oceanic slabs under the continents. This caused an anomalously high activity of suprasubduction volcanism;

(c) The minimum of the oceanic crust accretion in the interval of 80–70 Ma was reflected in a sudden decrease in the sea-floor spreading rate and oceanic crust subduction and, consequently, in weaker suprasubduction volcanism. The slower convergence of oceanic and continental plates was responsible for the crustal extension and intraplate basaltoid eruptions in separate parts of continental margins;

(d) The interdependent evolution of the Cretaceous continental margins and oceans was further complicated by the different dynamics of crustal generation in different oceans (Zonenshain and Kuz'min, 1992). For example, the sudden reduction of the spreading rate in the Pacific during the Campanian–Maestrichtian was synchronous with the crustal accretion maximum in the Atlantic that caused the Laramide tectogenesis, whose intesity was much lesser than that of other Cretaceous orogenic phases, and the Laramide orogenic belts in continental margins include some areas of crustal extension;

(3) The Cretaceous impulse of oceanic crust development corresponds to the extraordinary continental crust accretion. This accretion, depending on the rate of oceanic crust generation, resulted either in both the greater area and thickness of continental margins (due to accretion–collision and obduction processes) or just in the growth of their crustal thickness owing to accumulation of large masses of suprasubduction volcanites;

(4) The peak of Cretaceous climate warming during the interval of 120–80 Ma was correlated with the Early–Middle Cretaceous reorganization of lithospheric plates, with the appearance of an equatorial current, and with the extraction of extreme amounts of CO_2 from magmatic products of different types. It can be presumed that a large share of CO_2 increment in the atmosphere was related to the degazation of melts in the global systems of volcanic belts, whose peak activity coincides with the Albian–early Campanian temperature maximum (Huber *et al.*, 1995). Peculiar features of development of Cretaceous oceans and their continental margins were the main causes of climatic warming and great productivity of flora and fauna in the interval of 120–80 Ma, when they ultimately produced extremely large accumulations of coals, black shales, oil, and gas (Larson, 1991a, 1991b);

(5) The activity of plate tectonic events, including the evolution of active continental margins, was genetically related with the intensity of the lower mantle upwelling in the form of plumes generated at the core–mantle boundary (Larson and Olson, 1991). Stages of reorganization (including the Pangea break-up) and accelerated movements of lithospheric plates together with more active orogeny and magmatism corresponded to episodes of intense lower mantle upwelling, whereas decrease in plume activity was reflected in the slowing down of processes at the lithospheric level;

The time coincidence of activity maxima of synchronous mantle plumes caused the maximum intensity and global occurrence of events at the lithospheric level, e.g., in the interval of 120–80 Ma. On the other hand, the phase discrepancy of plume impulses was expressed in the development of tectonic processes of different types, e.g., in the interval of 80–50 Ma;

(6) Against the background of the permanently functioning Phanerozoic plumes of various degrees of intensity, the interval of 120–80 Ma stands out as the period of superplumes (Larson, 1991a). Its distinguishing features include extraordinary, even catastrophic events of all ranks including the lithospheric one. The break-up of Pangea and extremely intensive accretion of oceanic and continental crust at that stage coincided with the paroxysm of intraplate magmatism and with the period of cessation of magnetic field reversals;

(7) Judging from the consequences at the lithospheric level, two superplumes, the Atlantic–African the and Pacific, were active in the Cretaceous time. They differed in intensity, dynamics, and duration. The accretion rates of oceanic crust in the oceans show a significantly higher activity of the Pacific superplume compared with that of the Atlantic–African one. It is suggested that precisely this difference in intensities of these superplumes (the function of which, by the way, is the spreading rate) determined the differences in the character of boundaries with continental margins in the Pacific and Atlantic oceans: they were convergent in the former and passive in the latter.

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Correlation of the Middle Triassic Deposits in the North Caspian and Germanic Basins

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Abstract—The analyzed distribution ranges of tetrapods, ostracods, megaspores, and other fossils were used, with due regard for the facial dependence and phylomorphogenetic lineage, to resolve correlation problems concerning Triassic sections of the Germanic and North Caspian basins and southern Cis-Urals, and also to correlate them with the general stratigraphic scale of the Triassic. The Inder and Donguz formations of the North Caspian basin and Cis-Urals are proved to be early–middle Anisian in age. They correspond to the upper R_t (upper Buntsandstein) and lower Muschelkalk (lower–middle Wellenkalk) of the Germanic basin. The Sarpa and Masteksai formations of the North Caspian basin, along with the Yushatyr and Bukobai formations and a lower part of the Surakai Formation, are dated back to the late Anisian–Ladinian and correlated with the uppermost middle–upper Muschelkalk (its Anhydrite Group and Hauptmuschelkalk) and Lettenkeuper of the Germanic basin. A hiatus separating the Inder Formation from the Sarpa Formation and the Kiil Formation from the Masteksai Formation in the North Caspian region is coeval with that separating the Donguz and Yushatyr formations of the southern Cis-Urals. It appears to be a consequence of regression also recorded in the middle Muschelkalk (middle Illyrian) of Germany. In the North Caspian basin and Cis-Urals, the hiatus ranges from the middle-terminal Pelsonian to the mid-Illyrian.

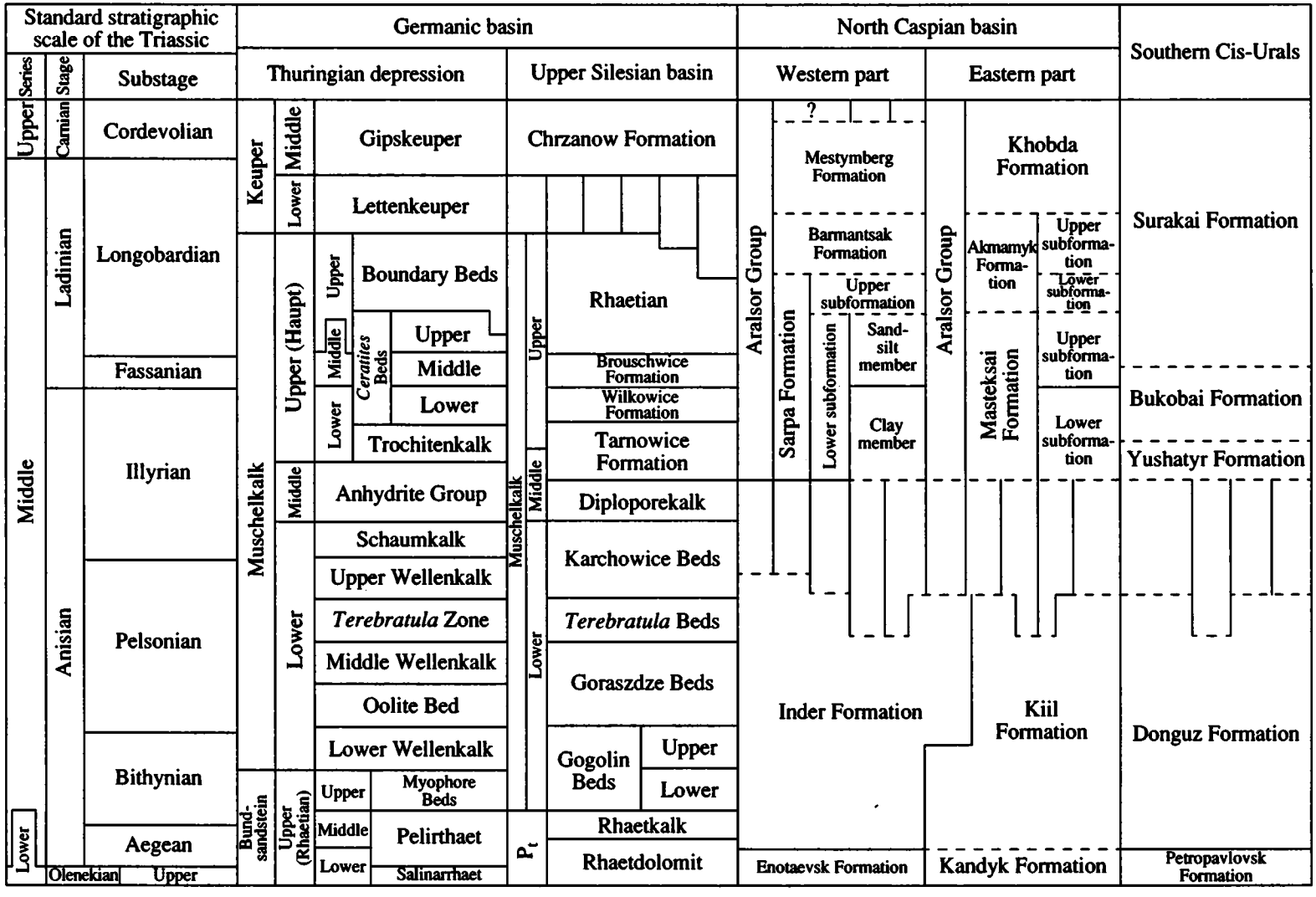
Key words: *correlation, Middle Triassic, ostracods, tetrapods, North Caspian and Germanic basins.*

In a series of publications of 1975–1981 (Movshovich and Kozur, 1975; Movshovich *et al.*, 1979; Movshovich, 1977, 1980, 1981), my colleagues and I elaborated the Triassic stratigraphic scheme for the southwestern Caspian lowland (Movshovich) and also revised the Triassic stratigraphy of the Germanic basin and correlated it (on the basis of ostracods, conodonts, charophytes, megaspores, and other fossils) with the Alpine and North Caspian Triassic (Kozur). The North Caspian paleontological materials used in these works were from my well-documented collection. For the last 15–20 years, our principal concept has not been changed, though it has been corrected to a certain extent.[1] For instance, the basal *Gemmanella* Beds of the Aralsor Group (formerly Formation) discordantly overlying the Inder Formation were classified as the Sarpa Formation (Movshovich, 1981). Moreover, it was shown that this formation is correlative in the east of the North Caspian basin not only with the Masteksai Formation, as is accepted after V.V. Lipatova in the unified stratigraphic scheme, but also with the lower part of the Akmamyk Formation (Movshovich, 1981, 1994; Shelekhova *et al.*, 1988). In addition, the Sarpa Formation was divided into two subformations (lower and upper *Gemmanella* Beds), and the lower of these units was shown to consist of clay and sand–silt members comprising respectively two lower and one upper ostracod zone of the *Gemmanella* Beds.

The Sarpa Formation is discordantly overlain by the Middle Triassic Barmantsak Formation (Table), an equivalent of the upper subformation of the Akmamyk Formation (Movshovich, 1994). Southeast of the stratotype of the Sarpa and Barmantsak formations, there is a 270- to 330-m-thick sequence of variegated sand and clay beds bearing a charophyte assemblage of the middle–upper Triassic and discordantly overlying the Barmantsak Formation. I classified this sequence, presumably an equivalent of the Khobda Formation of the North Caspian region, as the new Mestymberg Formation with the stratotype recovered by Well 2 (depth interval of 1914–2153 m) and parastratotypes recovered by wells 9 (depth interval of 1914–2122 m) and 4 (depth interval of 1937–2245 m) of the Mestymberg locality in the Kasatkinskaya prospect area, Priozernyi district, Kalmyk Republic.

Factual data on these stratigraphic subdivisions have already been published (Ochev, 1966; Tverdokhlebov, 1967; Lipatova *et al.*, 1972; Kozur, 1974a, 1974b; Ochev and Smagin, 1974; Movshovich and Kozur, 1975; Tuzhikova, 1975, 1979; Kozur and Movshovich,

[1] Dr. H. Kozur, who did not find an opportunity to work jointly on the manuscript, sees no need to be my coauthor, although he shares the conclusions presented here. I am very grateful to him for his helpful comments on certain points, which are carefully taken into account in this study.



Correlation scheme for the Middle Triassic deposits in the Germanic (modified after Kozur, 1974a, and Tröger *et al.*, 1984), North Caspian, and southern Cis-Urals basins.

1976; Kukhtinov, 1976; Tverdokhlebov and Makarova, 1976; Minikh, 1977; Movshovich, 1977, 1980, 1981, 1994; Movshovich *et al.*, 1979; Shelekhova *et al.*, 1988; Shishkin and Ochev, 1992), and I will consider below only that information that is necessary for discussion.

Recently, Shishkin and Ochev (1992), who considered the age of local subdivisions on the basis of the Middle Triassic tetrapod faunas of *Eryosuchus* and *Mastodonsaurus* genera, criticized our scheme of interregional correlation. They suspect that Triassic ostracods cannot be used at present in the interregional correlation (Shishkin and Ochev, 1992, p. 34), because there is a disagreement between the distribution of tetrapod remains in their interpretation and our ostracod-based correlation scheme. In fact, as it was formulated by Meyen (1982, p. 96), we must decide here "which one of several homotaxial successions should be taken for synchronous in different sections."

Shishkin and Ochev (1992, p. 34) point to a personal communication of Kukhtinov, who doubted the possibility of correlating Triassic deposits of Germany and the North Caspian region on the basis of ostracods, because respective assemblages of these fossils display "a distinction rather than a similarity." However, Kukhtinov personally (1976) and other specialists on ostracods (e.g., N.N. Starozhilova and A.G. Shleifer) used precisely these fossils in order to correlate Triassic deposits inside the North Caspian basin and between this and Germanic basins despite all the known differences in composition of respective ostracod assemblages (this kind of distinction is inevitable for any fauna and flora, since facial conditions of sedimentation are variable even in a single basin).

According to Kozur (1969, 1971, 1972), the ostracod-based correlation is rather reliable, when the phylogenetic level of fauna evolution and the basin salinity are taken into account. In this case, the appearance of one or another taxon can be recognized well below its first occurrence level in the originally studied sections.

Shishkin and Ochev cited Kukhtonov's opinion that *Gemmanella* forms appear beginning from the Ladinian Stage of Triassic deposits in Poland and Bulgaria. With respect to the last country, this is correct and may be a consequence of the fact that brackish-water deposits, where gemmanells should be concentrated, are missing in the upper Anisian marine sections of the region. However, such a statement is incorrect with respect to Poland. As shown by Kozur (1974a), gemmanells are very abundant in the uppermost beds of the middle Muschelkalk, which was referred to the basal upper Illyrian (Tröger *et al.*, 1984) in both northern Germany and northern Poland.

In the upper part of the middle–mid-upper Muschelkalk (upper Illyrian–Fassanian) of the Germanic basin, gemmanells coexist with abundant *Glorianella efforta* only in brackish-water, but not in the normal marine, deposits. The last form is characteristic of the Mastek-sai Formation and lower Sarpa Subformation, and this feature distinguishes ostracod assemblages of these units from the ostracod fauna of the Inder Formation. The last fauna is rich in smaller primitive (resembling *Lutkevichinella* species) glorianells, the precursors of these forms lacking peculiar marginal structures on the anterior extremities of both valves (Kozur, 1973). Similar forms are present in the lower Muschelkalk of the Germanic basin together with Bithynian–Pelsonian ammonoids and conodonts (Kozur, 1974a).

Even if Kukhtinov (1976) is right in stating that gemmanells appeared beginning from the Ladinian Stage, this statement is to be simultaneously considered as discrediting the idea of Shishkin and Ochev (1992, p. 29) that the Bukobai Formation, enclosing the *Mastodonsaurus* fauna and reasonably compatible with the *Gemmanella*-bearing Masteksai Formation, cannot be correlated with the lower Ladinian beds of the upper Muschelkalk in central Europe. Anyway, Kukhtinov did not study personally the Triassic ostracod collections from Germany, in contrast to Kozur (1974b), who himself studied their assemblages in this country and in the North Caspian basin. Therefore, Kukhtinov's viewpoint on distribution patterns of Triassic ostracods in the Germanic basin does not appear to be the most correct.

Further, Shishkin and Ochev (1992, p. 34) are skeptical with respect to our detailed correlation of ostracod assemblages from central and eastern Europe, because there is a disagreement between specialists on ostracods concerning the correlation of Triassic deposits in the Caspian depression. However, they ignore the fact that our inferences were based on the comparison of large phylomorphogenetic successions in the ostracod evolution, which were established by Kozur for different genera and groups of species but not on a simple correlation of assemblages on the basis of common elements. It is also important that, by this procedure, we compared identical or close species occupying the same place in the evolutionary succession.

The real abundance of ostracod species in various parts of the Triassic sections is sufficient for the detailed correlation. For instance, in the Triassic beds of the Germanic basin, Kozur confidently identified all the ostracod genera and many important species known in the North Caspian region, including some unusual forms, such as *Casachstanella* species.

Shishkin and Ochev, if they want to be logical, should identify the lower part of the Inder Formation stratotype, in accordance with principles they declared with respect to tetrapod remains, as an equivalent of the Donguz Formation, but instead they consider these remains as fossils of the Elton Horizon distinguished on the basis of ostracod fauna alone and unrecorded in the tetrapod locality of the Donguz Formation. I have already considered the problem of the poor validity of this horizon (Movshovich, 1981) and may omit it here.

In addition, I feel it necessary to recall here Ochev's opinion that large intervals of Triassic sections dated on

the basis of ostracods "are fairly well traceable" (Ochev and Smagin, 1974, p. 80). At the same time, the tetrapod remains (especially those that are determinable at the generic level at least) are much less abundant even in the well-exposed sections, though they are of high stratigraphic significance. It is also important to mention how Ochev (1979, p. 13) himself formulated the application limits of tetrapod fauna in correlation: "When we consider the detailed stratigraphy, i.e., the stage (group) subdivisions based on fauna groupings of the second order, all advantages of fossil tetrapods disappear. This fact is not always correctly understood."

One should also bear in mind another trivial rule that any fauna (faunal grouping) has a certain stratigraphic range. Therefore, stratigraphic levels of this fauna in any section indicate that fossils belong to this interval, yet they do not prove that the entire interval is present in a given case, if we have not recorded a transition from one fauna to another.

In the case under consideration, there is a clear discordance between the *Eryosuchus* and *Mastodonsaurus* faunas of the Donguz and Yushatyr formations. These faunas replaced one another during the time interval of the hiatus,[2] and in this case we deal not with changes in phylogenetic successions of the subfamily rank but with the change of different tetrapod families. Therefore, bone remains of a given fauna group, e.g., mastodonsaurs or plagiosaurs from different sections far remote one from another, do not suggest synchronism of bone-bearing beds but point to their position inside one stratigraphic interval.

For instance, Shishkin and Ochev consider the Kara-Bala-Kantemir tetrapod locality near Inder Lake as coeval with that of the Bukobai Formation, and I agree with this despite the fact that progressive plagiosaurs are rare in the southern Cis-Urals, whereas progressive *Plagiosternum* forms are missing near Inder Lake. Moreover, species of both groups, though different from their counterparts in the North Caspian and southern Cis-Urals regions, coexist with another genus of mastodonsauroid tetrapods in the Lettenkeuper of the Germanic basin. Accordingly, such tetrapod remains can be correlated only by levels of their phylogenetic evolution, and their age can be evaluated with a considerable error. Anyway, even data on tetrapod distribution cannot substantiate the strict age similarity of their localities in Germany and in the North Caspian and southern Cis-Urals regions.

There are several finds of prime significance for the age evaluation of the Donguz Formation: (1) a fragment of the lower jaw of *Eryosuchus(?) silesiacus* was recovered from the lower Wellenkalk of Silesia (Ochev, 1966, pp. 137 and 158; Ochev, 1968, p. 123; Shishkin and Ochev, 1992, p. 29); (2) the lower jaw of *Mentoposuarus walteri* resembling the *Eryosuchus* form was also collected from the lower Wellenkalk of Thuringia (Ochev, 1966, pp. 137 and 158); (3) poorly preserved fragments of capitosauroid bones assigned to the *Eryosuchus* fauna are known from dolomitic sandstone and tile clay beds (Rueau Sandstone of the Wellenkalk) exposed in the Mado River valley south of Nancy, Lorraine (Shishkin and Ochev, 1992, p. 29); (4) cranium fragments of *Aranetsia improvisa* from basal beds of the Krasnyi Kamen' Formation and ribs of a large *Plagiosternum* species from basal beds of the upper Lestanshor Subformation of the Pechora basin (Novikov and Shishkin, 1992) associate in these units with the Anisian (probably early Anisian) miospores (Il'ina and Novikov, 1991). All listed units, especially in the Germanic basin, correspond to the upper Bithynian and, partly, Pelsonian intervals of the middle Anisian, but not to the Ladinian Stage, as was postulated by Shishkin and Ochev.

These researchers rightly believe that the lower age limit of the middle–upper portion of the Donguz Formation (its lower portion lacks fossils) corresponds to the lower distribution limit of *Heptasaurus–Eocyclotosaurus* fauna. Its guide species *Heptasaurus cappelensis* (a primitive mastodonsauroid), *Stenotosaurus semiclausus,* and *Eocyclotosaurus woschmidti* are from that part of the R_t (upper Buntsandstein) in the Schwarzwald that seems to be correlative with basal (Aegean) beds of the Anisian. Certain species of this fauna (*H. waslensis* and *E. lehmani*) also occur in the Woltz Sandstone of Lorraine (an equivalent of the upper R_t corresponding to the lower part of the Bithynian). Consequently, the Aegean–lower Bithynian range of *Heptasaurus–Eocyclotosaurus* fauna of central Europe may well be correlative with the lower part of the Donguz Formation spanning an interval between its *Parotosuchus* and *Eryosuchus* faunas.

In addition, the middle–upper portion of the Donguz Formation includes primitive *Plagioscutum ochevi* and *Plagiosternum paraboliceps.* This portion appears to be spanning the uppermost Bithynian and Pelsonian. Thus, it is likely that there was a time interval of coexistence of *Heptasaurus–Eocyclotosaurus* and *Eryosuchus* faunas.

The upper part of the Hauptmuschelkalk (lower Langobardian) contains remains of large capitosauroid tetrapods that used to be referred to the *Mastodonsaurus* group, although in the opinion of Shishkin (1987, p. 5) "their generic affinity is rather unclear." Nevertheless, the conclusions of H. Mayer and J. Corrua, who suggested the presence of *Mastodonsaurus* remains at this level, seem to be correct, even if the determinations of Ochev are conditional (Shishkin and Ochev, 1992, p. 32). Shishkin and Ochev (ibid., p. 31) suggested that vertebrae from these beds belong to a capitosauroid form similar to *Eryosuchus.* The beds also yielded bones of

[2] If, as stated, "the fragment of the lower jawbone described by F. Broiley (1906) from the Middle Triassic of the Beirichian Alps and displaying the anteroacuminate Meckel's foramen typical of *Eryosuchus* forms" (Ochev, 1966, p. 137) really represents this genus, one should keep in mind that the fragment was recovered from the lower Carnian Reibl Beds of Bavaria. This actually means coexistence of *Eryosuchus* and *Mastodonsaurus* faunas.

ancient *Plagiosternum granulosum* and *Plagiosuchus pustuloglomeratus,* which differ from respective taxa of the Donguz Formation.

In my opinion, it is impossible to consider this fauna as identical to the *Eryosuchus* fauna of the Donguz Formation, and also to exclude its affinity with the *Mastodonsaurus* fauna, whose guide forms are either depressed or poorly preserved. Taking into account all the data, especially the hiatus between the Donguz and Yushatyr formations (Table), distinguished by Ochev (1968, p. 122), I believe that the first unit partially or completely spans the lower–middle interval of the Anisian Stage (Movshovich, 1980).

The lower part of the Inder Formation stratotype yielded separate bones of *Plagioscutum ochevi* (Shishkin, 1987, p. 9), vertebrae of "capitosauroid tetrapods similar in size and ossification degree with vertebrae of *Eryosuchus* from the Donguz Formation of the southern Cis-Urals" (Ochev and Smagin, 1974, p. 79), and the dental plate of *Ceratodus bucobaensis* (Minikh, 1977, p. 27). These fossils do not disprove the idea that the Donguz and Inder formations are coeval and overlie Petropavlovsk and Enotaevsk formations of the Olenekian Stage (table).

Ostracods are almost missing from the Donguz Formation but abundant in the Inder Formation. This fauna provides grounds to correlate the Inder Formation with the middle–upper R_t and lower Muschelkalk (Aegean–Pelsonian) of the Germanic basin. The upper half of the lower Muschelkalk yielded here abundant Pelsonian ammonoids of the *Balatonites* group associated with conodonts (*Nicorella kockelli, N. germanica,* and *Neogondolella bulgarica*) and brachiopods (*Decurtella decurtata*).

A hiatus established in the North Caspian and southern Cis-Urals regions, where it separates the Inder, Kiil, and Donguz formations, from the Sarpa, Masteksai, and Yushatyr formations, respectively, is a consequence of regression, whose marks are also recorded in the middle Muschelkalk (mid-Illyrian) of the Germanic basin. In the North Caspian and Cis-Urals sections, the hiatus spans an interval of the stratigraphic scale beginning from the middle–upper Pelsonian (the concrete level depends on the irregular depth of incision into the top surface of the Inder, Kiil, and Donguz formations) to the mid-Illyrian (table).

Shishkin and Ochev (1992, p. 34) argued against my inference that lower beds of the Inder Formation are early Anisian in age. This formation was correlated with the R_t of Poland (Shelekhova, *et al.,* 1988) on the basis of palynological data, and my opponents suggest the last unit to be not older than the middle Anisian. However, Kozur (1974a) has shown the R_t of Poland to correspond better to the interval of Aegean–lower Bithynian, whereas its basal beds are correlative with the upper Olenekian (Table), and his data support my opinion. I also believe that *Dadocrinus gracilis,* the guide fossil of the lower Bithynian reported from the Gogolin Beds of Upper Silesia resting on the R_t, completely discredits the opinion that in Poland the latter unit is "not older than the middle Anisian" (Shishkin and Ochev, 1992, p. 34).

With respect to the conclusion about the early–middle Anisian age interval of both the Donguz and Inder formations, my opponents consider it as a crude mistake of mine. Nonetheless, at the time of my works, Ochev himself was sure that the Donguz Formation is an equivalent of the upper Buntsandstein and lower Muschelkalk; i.e., he correlated it with the lower half of the middle Triassic in contrast to his recent opinion that this formation (or its greater upper portion) is correlative with the lower and basal upper Ladinian.

According to Shishkin and Ochev, the Donguz and Inder formations span an abnormal stratigraphic interval (the entire Anisian Stage together with the main part of the Ladinian Stage). Considering that the Triassic sections in question do not display considerable breaks in sediment accumulation and traces of uncompensated sedimentation, this assumption seems to be especially improbable, when we compare the lithology and thickness of both units with the same characteristics of younger and older formations.

Now I would like to discuss available age data on the Kara-Bala-Kantemir locality of tetrapod remains situated at the crest of the Kok-Tau Ridge near the eastern coast of Inder Lake. In 1984, V.R. Lozovsky (personal communication) studied this locality, when it was much better stripped by bulldozer than before. He concluded that bone-bearing clay beds with sandstone interlayers should be referred here to the Masteksai Formation, because the underlying clay of the Inder Formation with thin limestone interlayers contains only rare indeterminable bone remains. Shishkin (1987, p. 9) also reported that green and variegated bone-bearing clay beds rest with incision marks on the barren black clay.

Twenty years ago (Movshovich, 1977), I erroneously assigned the bone-bearing clay beds to the uppermost portion of the Inder Formation. The same mistake of Lipatova *et al.,* (1972) and Ochev and Smagin (1974) resulted from the position of the beds in the Kara-Bala-Kantemir locality close to the boundary between the Inder and Masteksai formations, and the originally poor state of this exposure was unfavorable for the precise positioning of the bone-bearing level.

The greenish gray and variegated clays of the Masteksai Formation yielded bones of *"Mastodonsaurus" torvus* and less abundant *Plagioscutum caspiense.* In the clay samples, which were collected here by Lozovsky, M.N. Shelekhova distinguished an assemblage of miospores identical to that typical of the Masteksai and Bukobai formations. In the locality under consideration, the Masteksai clay fills incisions in the topmost dark gray to black clay beds of the Inder Formation, whose thin limestone interlayers contain an assemblage

of the lower Pelsonian ostracods (Movshovich and Kozur, 1975, pp. 109–110; Movshovich, 1977, p. 99).

I reported about all these new data correcting my previous concept (Movshovich, 1980, 1981) at the plenary session of the Regional Interdepartmental Stratigraphic Commission on the Central and Southern Russian Platform, November 1991, Saratov.

In the opinion of Tuzhikova (1975), the miospore assemblage from the basal (bone-bearing) beds of the Bukobai Formation stratotype is older than that typical of the Lettenkeuper in central Europe. The Illyrian age of the lower part of the Masteksai Formation is substantiated by characteristic ostracods and miospores recovered from the clay member of the lower Sarpa Subformation (60 m above its base, see table) penetrated by Well 5-c (depth interval of 1114–1120 m) in the Vladimirovo region (Kozur and Movshovich, 1976; Movshovich *et al.*, 1979). This ostracod assemblage was described earlier (Movshovich *et al.*, 1979), and I omit its characteristics here.

The age of the Bukobai Formation is indirectly determinable on the basis of the ostracod assemblage from Well C-36 of the Gir'yalskaya Geological Survey Party (depth level of 108 m), where it is confined to basal beds of the Surakai Formation composed of gray and greenish gray, fine- to medium-grained feldspar–quartz silt and sand overlying the Bukobai Formation (Movshovich, 1980). A.N. Baidova, the chief geologist of the party who led the geological survey of the Novocherkasskii basin in the Cis-Urals (the left-hand side of the Ural River northeast of the Novocherkasskii settlement) by studying natural exposures, prospect holes, and drilling sites, informed me (personal communication) that the silt–sand member of the Surakai Formation (Well C-36, the eastern part of the basin) rests here on the basal clay bed overlying the variegated (mostly red) Bukobai Formation. Other researchers (Garyainov and Shatkinskaya, 1966; Tverdokhlebov, 1967) also reported about these characteristic Surakai beds of the Novocherkasskii basin.

The ostracod assemblage from this structure includes *Bisulcocypris rubachae,* a primitive form related to the older species *Lutkevichinella* (*Cytherissinella*) *sokolovae*, and *Gemmanella meyeni* related with its precursor *Gemmanella* (*Praegemmanella*) *subtilis;* and associated with *G.* (*Neogemmanella*) *oertlii*, *Glorianella efforta*, and *Blomella transita* (a form transitional to the older species *B. sulcata*).

This assemblage characteristic of the silt–sand member of the lower Sarpa Subformation (Table) is similar in its features to the transitional Illyrian–Fassanian ostracod assemblages. Accordingly, it would be more correct to place the Bukobai–Surakai boundary inside the basal portion of the Fassanian, but not at the base of this unit, as I thought before (Movshovich, 1980).

This position of the boundary is concordant with data on the identical ostracod assemblages recovered from the upper part of the Masteksai Formation at the Krasnoyar Cupola located in the east of the North Caspian region and from the lower part of the Surakai Formation (Starozhilova and Shelekhova, 1987). Shelekhova, who studied palynology of the Masteksai Formation at the Krasnoyar Cupola and other localities, suggested the early Ladinian age for its upper portion.

Shishkin and Ochev (1992, p. 34) also insist that data of Tverdokhlebov and Makarova (1976) on the Bukobai Formation stratotype invalidate my age determination for the Surakai Formation of the Novocherkasskii basin. In fact, these researchers studied the stratotype area of the Surakai Formation, and earlier I published my negative reaction to their conclusions (Movshovich, 1980).

It is more important to mention here the opinion of Tuzhikova (1979, p. 27), who studied the same area. As she observed, the aforementioned work of Tverdokhlebov and Makarova (1976) "simply demonstrates to readers how the gray-colored rocks of different stratigraphic units of the southern Cis-Urals were ascribed altogether to the Surakai Formation just on the basis of visual recognition of their color in various exposures without a clear differentiated insight into their lithology. Afterward, the Surakai Formation, thus transformed into a collecting unit for all gray-colored deposits of the southern Cis-Urals, was declared invalid because of its "diachronism," and its rank was defined as the Surakai facies. This answer to the problem of the Surakai Formation can be hardly considered as correct."

Supporting the judgment of Tverdokhlebov and Makarova that the "Surakai facies" replaces the Donguz Formation of the Surakai basin, Shishkin and Ochev overlooked a clear conflict between this assumption and the data of many researchers, including Garyainov, Ochev, and Tverdokhlebov, who insist that there is a regional hiatus separating the Donguz Formation from the Yushatyr, Bukobai, and Surakai formations overlying the former.

Studying the miospore assemblage from the uppermost beds of the Bukobai Formation stratotype, Tuzhikova (1975) arrived at the conclusion that it is similar to the assemblage typical of *Gemmanella* Beds (Masteksai Formation) of the North Caspian region. She also concluded that another miospore assemblage from the Surakai Formation stratotype is younger and suggests the early Carnian age of the host rocks (Tuzhikova, 1979). However, in the last case, Tuzhikova (ibid., p. 30) remarked that lower parts of this formation appear to be inaccessible in both the drill holes and natural outcrops of the Surakai basin. I completely agree with her dating that assumes the Late Triassic age for the upper part of the Surakai Formation (Movshovich, 1980).

The late Illyrian–Fassanian age of the lower Sarpa Subformation is substantiated by ostracods and miospores. Moreover, this dating is logically compatible with the Longobardian age of the upper Sarpa Sub-

formation. The last conclusion is based on rare ostracod remains (Movshovich *et al.*, 1979) and miospores (Kozur and Movshovich, 1976) recovered from the upper half of the upper Sarpa Subformantion by wells Sadovaya 1-p (depth intervals of 1844.3–1905.5 and 1920–1924 m) and Tsaryn-6 (depth interval of 2440–2450 m) and found to be corresponding to the Longobardian fossils of Germany. The Middle Triassic ostracods and miospores from the Barmantsak and Berg formations are poorly studied, and I omit here their analysis.

Assuming the Late Triassic age of the Surakai Formation, Shishkin and Ochev (1992, pp. 31, 32) disregard the data of Dobruskina (1982), who attributed phytological remains from this and the underlying Bukobai Formation to the single sciophilous flora of the Ladinian–Carnian age also exemplifying floral remains of the Kurashasai and Kuraily formations. These data do not disprove the Ladinian age of the lower Surakai deposits.

As mentioned above, the Late Triassic (Carnian) age of the upper part of the Surakai Formation is substantiated by miospores (Tuzhikova, 1979). Judging from megaspores, the Kurashasai Formation is of the same age (Movshovich, 1980). Some geologists (Tverdokhlebov and Makarova, 1976) believe that miospores from the youngest Surakai beds recovered by wells are Rhaetian or even Rhaetian–Liassic in age. At the same time, tetrapod remains systematically occurring in the Upper Triassic sections of western Europe are still unknown from coeval deposits of eastern Europe.

The precise origin time of the *Mastodonsaurus* fauna is unknown. It is true that some capitosauroid remains do not belong to this fauna (Shishkin and Ochev, 1992, p. 32), as I formerly thought following an earlier assumption of Ochev. Nonetheless, there is no doubt that most primitive mastodonsauroids (*Heptasaurus*) already existed at the beginning of the Anisian. Moreover, describing a bone of *Mastodonsaurus* sp. from the middle part of the Donguz Formation, Ochev (1972, pp. 196–197) observed that "it is indistinguishable from the respective bone of *M. torvus* Konzh. This find indicates that the *Mastodonsaurus* genus existed at the Donguz time, even if it was very rare."

It is clear that the *Mastodonsaurus* fauna became rather widespread before the accumulation time of the Yushatyr Formation, presumably during the hiatus time separating this unit from the Donguz Formation and caused by the mid-Illyrian regression. It is also noteworthy that Tverdokhlebov and Makarova (1976, p. 58) believe that this fauna appeared at the beginning, but not in the second half, of the Ladinian Age, although Shishkin and Ochev (1992, p. 29) cited precisely this work to support their own speculations. In a similar manner, they referred to conclusions of Tuzhikova despite the fact that miospores from the basal (bone-bearing) beds of the Bukobai Formation are, in her opinion (Tuzhikova, 1975, p. 6), older than those from the Lettenkeuper.

Thus, Shishkin and Ochev (1992), who contributed much to the paleontological characterization of deposits in question, do not disprove in principle my correlation scheme of the Middle Triassic stratigraphy in the North Caspian and Germanic basins. I consider it useful to discuss the problems concerning this issue. Such a discussion allowed me to show that, accepting the lower–middle Anisian (Aegean–Pelsonian) interval for the coeval Inder and Donguz formations, we have an agreement with the ostracod dating and actually avoid conflict with data on the tetrapod distribution. Localities of the *Mastodonsaurus* fauna in the North Caspian region (Masteksai Formation) and southern Cis-Urals (Yushatyr and Bukobai formations) appear to be older than respective localities in the Lettenkeuper of central Europe. This is also evident from a certain difference in their specific and, to a lesser extent, generic compositions.

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Upper Cretaceous Ammonite Zonation of Eastern Central Asia

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Abstract—This paper presents a new detailed ammonite-based biostratigraphic zonation of the Upper Cretaceous in eastern Central Asia and substantially refines the previous scales. Two units were established in the lower Cenomanian: the *Mediasiceras beljakovae* and *Karamaites gaurdakense* beds. The middle Cenomanian comprises three units: the *Mediasiceras lenticulare* (Luppov) Beds, the *Kopetdagites aktaschensis* Zone, and the *Acanthoceras jukesbrownei* Zone. The upper Cenomanian includes the *Eucalycoceras pentagonum* and *Sciponiceras gracile* zones. In the lower Turonian, the *Watinoceras coloradoense* and *Thomasites koulabicus* zones are distinguished. The middle Turonian is divided into the *Spathites (Jeanrogericeras) reveliereanum* and *Colllignoniceras woolgari* zones. The upper Turonian corresponds to three units: the *Subprionocyclus neptuni* and *Hourcquia pacifica* zones and the *Coilopoceras gissarensis* beds. The *Stantonoceras guadalupae* and *Stantonoceras polyopsis* zones are recognized in the lower Santonian, and the local *Asiatostantonoceras tagamense* Zone is established in the upper Santonian. The lower Campanian corresponds to the single *Scaphites inflatus* Zone, whereas the upper Campanian is sudivided into the *Hopliplacenticeras marroti* and *Trachyscaphites pulcherrimus* zones. The *Nostoceras hyatti* and *Baculites knorrianum* zones are identified in the lower Maastrichtian. The lower boundary of the Maastrichtian is placed at a significantly lower level as compared with the previous scales. All the defined units are correlated to stage stratotypes and stratigraphic schemes of the adjacent regions.

Key words: *stratigraphy, Upper Cretaceous, ammonites, eastern Central Asia.*

The Upper Cretaceous deposits are widespread in eastern areas of Central Asia, e.g., in Tien Shan and its intermontane depressions. They are irregularly exposed in core and flank areas of almost all the anticlines of the Tajik depression and its framing structures, such as the southwestern outlets of the Gissar Ridge and southwestern Darvaz, in separate depressions of the Zeravshan–Gissar Mountains, along the southern slopes of the Alai and Transalai ridges, in the eastern and northern parts of the Fergana depression, and in the Naukat depression (figure). The Upper Cretaceous sequence is composed of sandy–clayey deposits with interbeds of carbonate rocks enriched in terrigenous material and detritus. In addition to marine facies, the sequence includes lagoonal and continental sediments. The share of the latter substantially increases eastward, and faunal assemblages simultaneously become progressively impoverished in the same direction. Species characteristic of the Mediterranean realm appear in addition to endemic forms.

Upper Cretaceous deposits of eastern Central Asia were investigated by many researchers (Mikhailovskii, 1914: Borneman, 1940; Simakov, 1952, 1953; Bobkov, 1961; Iljin, 1969; Poyarkova, 1969; Dzhalilov, 1971; and others).

For more than two decades (1963–1992), I studied these deposits, sampling ammonites and other fossils to compile detailed stratigraphic schemes. Based on the materials obtained recently, a more sophisticated and paleontologically well-substantiated stratigraphic scale of the Upper Cretaceous was elaborated for eastern Central Asia (Tables 1, 2). It significantly differs from the earlier zonation (Iljin, 1969).

The following stratigraphic units are introduced into this scale: beds with particular fauna and local and standard zones. Beds with fauna represent units of the local stratigraphic scheme. They represent deposits bearing certain species and assemblages, which distinguish the enclosing strata from underlying and overlying members. In this paper, units of this type are applied when distribution of corresponding deposits are restricted to one or several areas.

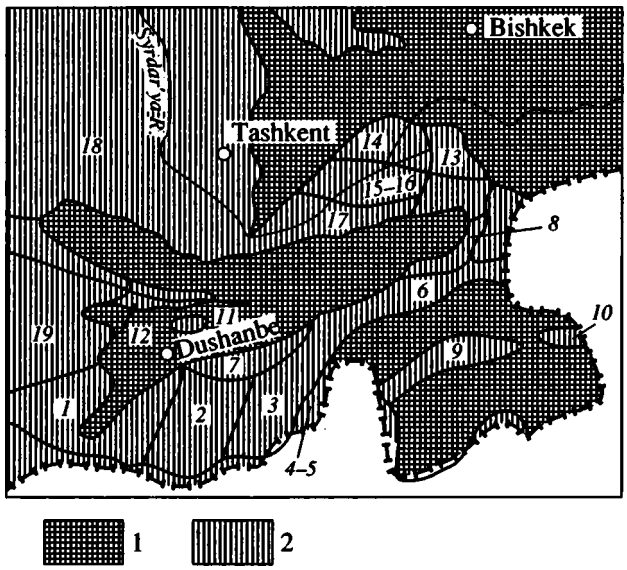
The local zone is a category of regional units valid for several regions (Bukhara–Tajik, Fergana).

The zone is a unit category of the standard stratigraphic scale. This unit is considered here as it is defined by Krymgol'ts (1980).

THE LOWER CENOMANIAN

Mediasiceras beljakovae Beds

This unit is established only in the Gaurdak area of the Bukhara–Tadjik region, where it corresponds to the



Schematic map of distribution of Upper Cretaceous deposits in Central Asia: (1) pre-Cretaceous deposits; (2) Cretaceous deposits. Areas (numbers in the scheme): (*1*) Gaurdak; (*2*) Kafirnigan; (*3*) Vakhsh; (*4*) Khozretishi; (*5*) Obiniou; (*6*) Transalai; (*7*) southern Gissar; (*8*) Alai; (*9*) Akbaital'; (*10*) Rangkul'; (*11*) Ravat; (*12*) Magian; (*13*) Karadar'ya; (*14*) Varzyk; (*15–16*) Naukat and Kan; (*17*) Isfarin; (*18*) Tashkent; (*19*) Bukhara.

Tyubegatan Formation. Beds conformably rest upon deposits of the *Stoliczkaia dispar* Zone (Vraconian).

Their ammonite assemblage includes *Mediasiceras beljakovae* Iljin and *Karamaites* aff. *mediasiaticum* (Luppov).

The index species of *Mediasiceras beljakovae* Beds was reported from the lower Cenomanian *Mantelliceras mantelli* Zone of the western Kopetdag (Atabekyan, 1966).

Karamaites gaurdakensis Zone

This zone is recognized in the Gaurdak area of the Bukhara–Tadjik region and corresponds to the Karakansai Formation.

Its ammonite assemblage consists of *Karamaites gaurdakensis* (Luppov), *K. gissarensis* Iljin, *K. asiaticum* (Iljin), *K. acutum* (Iljin), *K. mediasiaticum* (Luppov), *Mantelliceras* sp., and *Anaplacenticeras* sp.

The early Cenomanian age of the zone is confirmed by finds of *Mantelliceras* sp. Iljin (1969) detected this species in the lower Cenomanian deposits of the Badkhyz Mountains, where it occurs along with *Karamaites mediasiaticum* (Luppov), *Mantelliceras tuberculatum* Mant., *M.* cf. *hyatti* Spath, and *M. mantelli* Spath. The stratigraphic range of these species indicates the early Cenomanian age of the considered zone. The index species *Karamaites gaurdakensis* (Luppov) occurs in the lower Cenomanian *Submantelliceras saxbi* and *Mantelliceras mantelli* zones of the western Kopetdag (Atabekyan, 1966).

THE MIDDLE CENOMANIAN

Mediasiceras lenticulare Beds

Deposits of the unit in question are spread in the Gaurdak and Kafirnigan areas of the Bukhara–Tajik region and correspond to the lower part of the Gazdaganin Formation. Their ammonite assemblage consists of *Karamaites grossouvrei* (Semenov), *Mediasiceras lenticulare* Luppov, and *M. sagittalis* Iljin. The Cenomanian age of the unit is substantiated by the presence of *Karamaites grossouvrei* (Semenov) and *Mediasiceras lenticulare* Luppov species coexisting with *Turrilites costatus* Lam., *T. acutus* Passy, and *Euomphalloceras cunningtoni* (Sharpe) (Atabekyan, 1986).

The stratigraphic range of the above-listed ammonite species allows the beds in question to be correlated with the upper part of the *Euomphalloceras cunningtoni* Zone of the western Kopetdag (Atabekyan, 1986). Correlation through transitional areas shows that the unit probably corresponds to a part of the *Acanthoceras rhotomagense* Zone of western Europe (Kennedy, 1986).

Kopetdagites aktaschensis Zone

Deposits of this zone are widespread in the Bukhara–Tajik and Fergana regions, corresponding there to the lower part of the Gazdaganin Formation.

The ammonite assemblage from this zone includes *Kopetdagites aktaschensis* Iljin, *K. kopetdagensis* Iljin, *K.* aff. *aktaschensis* Iljin, *Acanthoceras* aff. *tapara* Wright, and *A. rhotomagense asiaticum* Iljin.

Table 1. Stratigraphic distribution of ammonites in the Upper Cretaceous of eastern Central Asia

Taxons	Cenomanian							Turonian								Coniacian		Santonian		Campanian				Lower Maastrichtian
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25
Mediasiceras beljiakovae Iljin	+																							
Karamaites aff. *mediasiaticum* (Luppov)	+																							
K. gaurdakense (Luppov)		+																						
K. gissarensis Iljin		+																						
K. asiaticum (Iljin)		+																						
K. acutum (Iljin)		+																						
K. mediasiaticum (Luppov)		+																						
Mantelliceras sp.		+																						
Karamaites grossovurei (Semenov)			+																					
Mediasiceras saggitalis Iljin			+																					
M. lenticulare (Luppov)			+																					
Kopetdagites aktaschense Iljin				+																				
K. kopetdagensis Iljin				+																				
Karamaites subtilis Iljin				+																				
Acanthoceras aff. *tapara* Wright				+																				
A. rhotomagense asiaticum Iljin				+																				
A. jukesbrownei (Spath)					+																			
A. sherborni (Spath)					+																			
Newboldiceras newboldi (Kossm)					+																			
Eucalycoceras pentagonum (J-Br)						+																		
Calycoceras cf. *naviculare* (Mantell)						+																		
Allocrioceras anulatum (Shumard)							+																	
Anisoceras plicatile (Sow.)							+																	
Euomphalloceras septemseriatum (Cragin)							+																	
Metoicoceras geslianinum (Orb.)							+																	
Borrisiakoceras orbiculatum (Stephenson)							+																	
B. mirabile Arkh.							+																	
Placenticeras cf. *memorishloenbachi* (L. et B.)							+																	
Sciponoceras gracile (Shumard)							+																	
S. baculoide (Mant.)							+																	
Vascoceras diartianum (Orb.)							+																	
Worthoceras vermiculum (Shumard)							+																	
Mammites nodosoides chivensis Arkh.								+																

Table 1. (Contd.)

1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25
Proplacenticeras kharesmense (Lahusen)								+																
Puzosiachivensis Arkh.								+																
Watinoceras coloradoense (Henderson)								+																
Beschtubeites cf. *alaiense* (Luppov)									+															
B. faasi Arkh.									+															
B. kysylcurrganense (Luppov)									+															
B. kutuzovae Iljin									+															
B. cf. *tenerum* Iljin									+															
Fagesia catinus (Mant.)									+															
Lewesiceras peramplum Mant.									+															
Proplacenticeras simakovi (Luppov)									+															
Nigericeras kleri (Atabekjan)									+															
Thomasites koulabicus Kler									+															
Lecointriceras fleurrisianum (Orb.)										+														
Nurekites costatus (Stankevich)										+														
N. braecostatus (Stankevich)										+														
N. bulgariensis Khakimov										+														
N. nurekensis Khakimov										+														
N. raricostatus (Stankevich)										+														
Neoptychites cephalotus Court.										+														
Prohauericeras goupilianum (Orb.)										+														
Sciponoceras bohemicum (Fritch.)										+														
Spathites (Jeangericeras) *revelierеanum* (Court.)										+														
Collignoniceras carolinum (Orb.)											+													
C. bakeri (Anderson)											+													
C. canthus (Sornay)											+													
C. turoniense (Sornay)											+													
C. woolgari (Mant.)											+													
Lecointriceras iljini Khakimov											+													
L. cf. *costatum* Kennedy, Wright, Hancock											+													

Table 1. (Contd.)

1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25
Prionocyclus cf. *hyatti* (Stanton)											+													
Pseudoforresteria charrochensis Khakimov											+													
Ps. babatagensis Khakimov											+													
Romaniceras cf. *kallesi* (Zazvorka)											+													
Lewesiceras sharpei (Spath)												+												
Prionocyclus cobbani Matsumoto												+												
Pr. cf. *wyomingensis* Meek												+												
Proplacenticeras cf. *arkhangelski* Iljin												+												
Pseudoforresteria ornata Iljin												+												
Ps. asiatica Khakimov												+												
Subprionocyclus branneri (Matsumoto)												+												
S. beschtubense Iljin												+												
S. cf. *normalis* (Anderson)												+												
S. neptuni (Qein.)												+												
S. inflatus Khakimov												+												
S. bravaisianus (Orb.)												+												
Coilopoceras gissarensis Iljin													+											
Placenticeras akrabatense Vinocurova													+											
Aktaschia akrabatense Iljin														+										
Houroquia pacifica Matsumoto														+										
H. aff. *hataii* Hashimoto														+										
H. cf. *mirabilis* Collignon														+										
Placenticeras kotzi Iljin														+										
Proplacenticeras proplanum Iljin														+	+									
Propl. cf. *orbignyanum* (Qeinitz)														+	+									
Prionocyclus cf. *quadratus* Cobban														+										
Reesidites minimus (Hausaka et Fukada)														+										
"Lewesiceras" asiaticum Iljin															+									
Peroniceras aff. *proteus* Mats.															+									
Placenticeras pitniakense Iljin															+									
Prionocycloceras guayabanum (Stein.)															+									

Table 1. (Contd.)

1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25
Pr. cf. *multicoststum* Coll.															+									
Pr. aff. *carvalhoi* (Howart)															+									
Eostantonoceras babatagensis Khakimov																+								
Placenticeras bobkovae Iljin																	+							
Pl. luppovi Iljin																	+							
Stantonoceras guadalupae asiaticum Iljin																	+							
S. kysylcumense Arkh.																	+							
S. polyopsis Dujardin																		+						
Asiatostantonoceras tagamense Iljin																			+					
Scaphites inflatus Roem.																				+				
Gissarites tagamense Iljin																				+				
G. kysylchense Iljin																				+				
H. marroti (Coq.)																					+			
H. vari (Schluter)																					+			
H. (*Lemfordiceras*) *rarecostatum* Khakimov																					+			
Trachyscaphites spiniger (Schluter)																					+	+		
T. gibbus (Schluter)																					+			
Didymoceras vancouverensis Gabb.																						+		
D. parastevensoni Atabekjan et Khakimov																						+		
Trachyscaphites pulcherrimus (Roemer)																						+		
Glyptoxoceras retrorsum (Schluter)																						+		
Baculites anceps Lam.																						+		
B. vertebralis Lam.																						+		
Nostoceras hyatti Stephenson																							+	
N. cf. *draconis* (Stephenson)																							+	
N. cf. *pozariskii* Blaszkiewicz																							+	
Neohamites subcompressus (Fofbes)																							+	
Baculites knorrianus Desmarest																								+

Note: Lower Cenomanian: (2) *Mediasiceras beljakovae* Beds, (3) *Karamaites guardakense* Zone, (4) *Mediasiceras lenticulare* Beds, (5) *Kopetdagites aktaschense* Zone, (6) *Acanthoceras jukesbrownei* Zone; upper Cenomanian: (7) *Eucalycoceras pentagonum* Zone, (8) *Sciponoceras gracile* Zone; lower Turonian: (9) *Watinoceras coloradoense* Zone, (10) *Thomasites koulabicus* Zone; middle Turonian: (11) *Spathites* (*Jeanrogericeras*) *revelierianum* Zone, (12) *Collignoniceras woolgari* Zone; upper Turonian: (13) *Subprionocyclus neptuni* Zone, (14) *Coilopoceras gissarensis* Beds, (15) *Hourcquia pacifica* Zone; Coniacian: (16) *Prionocycloceras guayabanum* Zone, (17) *Eostantonoceras babatagensis* Beds; lower Santonian: (18) local *Stantonoceras guadalupe asiaticum* Zone, (19) *Stantonoceras polyopsis* Zone; upper Santonian: (20) local *Asiatostantonoceras tagamense* Zone; lower Campanian: (21) *Scaphites inflatus* Zone; upper Campanian: (22) *Hoplitoplacenticeras marroti* Zone, (23) *Trachyscaphites pulcherrimus* Zone; lower Maastrichtian: (24) *Nostoceras hyatti* Zone, (25) *Baculites knorrianus*; upper Maastrichtian: (26) barren of ammonites.

Table 2. Correlation of the Upper Cretaceous stratigraphic schemes for the eastern Central Asia

Stage	Substage	Standard scale, Amedro *et. al.*, 1982; Schulz, 1979; Christensen, 1986	Western Central Asia, Atabekyan, 1986, 1991	Interior areas of Central Asia Iljin, 1969	Eastern Central Asia, *Resheniya...* 1971	Eastern Central Asia, this paper
Maastrichtian	upper	*Neobelemnella kazmirovensis* / *Sphenodiscus binkhorsti* *Belemnitella junior*	*Anapachydiscus fresvillensis*	*Belemnella arkhangelskii (Neobelemnella kazimirovensis)*	*Liostrea lehmanni* and *Biradiolites boljnanesis*	–
	lower	*Belemnella ossidentalis*: *Belemnella factigata*, *Belemnella cimbrica*, *Belemnella sumensis* / *Pachidiscus neubergicus* *Belemnella lanceolata*: *Belemnella obutsa*, *Belemnella dseudobutsa*, *Belemnella lanceolata* / *Sphenodiscus ubagshi*	*Diplomoseras cylyndroceum* *Pachyodiscus neubergicus* *Hauerisiras sulcatum*	*Belemnella lanceolata*		*Baculites knorrianus Zone*
Campanian	upper	*Bostrychoceras polyplocum* *Holitoplacenticeras marroti*	? *Bostrychoceras polyplocum* *Holitoplacenticeras coesfeldiense*	*Hoplitoplacenticeras marrioti*	*Hoplitoplacenticeras marrioti* and *Bostrychoceras polyplocum*	*Nostoceras hyatti Zone* *Trachscaphites pullacherimus Zone* *Hoplitoplacenticeras marroti Zone*
	lower	*Delawerella campaniensis* *Placenticeras bidorsatum*	*Eupachydiscus levyi* *Offaster pomeli*	*Scafites inflatus*	*Scafites inflatus*	*Scafites inflatus local Zone* –
Santonian	upper	*Placentiras polyopis* / *Eulophoceras austriacum*	*Marsupites testudinarus* *Uintacrinus socialis*	*Asiatostantonoceras tagamense*	*Asiatostantonoceras tagamense*	*Asiatostantonoceras tagamense local Zone*
	lower	*Texanites gallicus*	?	*Stantonoceras guadalupae asiaticum*	*Stantonoceras guadalupae asiaticum*	*Stantonoceras polyopsis Zone* *Stantonoceras guadalupae asiaticum local Zone*
Coniacian	upper	*Paratexanites seratomarginatus* *Gauthiericeras margae*		*Lewesiceras asiaticum*	*Lewesiceras asiaticum*	*Eostantonoceras Babatagensis Bads* *Hourcquia pacifica Zone*
	middle	*Peroniceras tridorsatum*	*Peroniceras tridorsatum*	*Barroisiceras haberfeleneri*	*Hourcquia akrabatense*	*Hourcquia pacifica Zone*
	upper	*Forresteria (Haleites) petracoriensis*	?			*Coilopoceras gissarensis Beds*
Turonian	upper	*Subprionocyelus neptuni* *Romaniceras deverianum* / *Collignoniceras woolgari*	*Subprionocyclus neptuni* ?	*Collignoniceras woolgari*	*Collignoniceras intermedium*: *Subprionocyclus cristatus*, *Collignoniceras woolgari*	*Subprionocyclus neotuni Zone*
	middle	*Romaniceras ornatissimum* *Romaniceras kallezl* *Kamerunoceres turoniense*	*Collignoniceras woolgari*			*Collignoniceras woolgari Zone* *Spathites (Ieanrogericeras) revelierianum Zone*
	lower	*Mammites nodosolides* *Watinoceras coloradoense*	*Mammites nodosolides* *Watinoceras coloradoense*	*Mammites nodosolides*	*Mammites nodosolides* and *Inoceramus labiatus*	*Thomasites koulabicas Zone* *Watinoceras coloradoense Zone*
Cenomanian	upper	*Neocardioceras juddi* *Metoicoceras geslinianum* *Calycoceras guerangeri*	? *Eucalycoceras pentagonum*	*Kopetdagites aktaschensis*	*Kopetdagites aktaschensis*	– *Sciponoceras grasile Zone* *Eucalycoceras pentagonum Zone*
	middle	*Alternathantoceras jukesbrownei* *Acanthoceras rhotomagense* / *Turilites acutus*, *Turilites costatus*	*Alternacanthoceras jukebrownei* *Pleciacanthoceras amphibolum* *Acanthoceras rhotomagense* *Guerangericeras confusum* *Gunningtoniceras gunningtoni*	*Eoradiolites kugitangesis*	*Eoradiolites kugitangesis*	*Acanthoceras rhotomagense Zone* *Kopetdagites aktaschensis Zone* –
	lower	*Mantelliceras dixoni* *Mantelliceras mantelli* / *Mantelliceras saxbii*, *Neostlingoceras rascarcitanesis*	*Mantelliceras dixoni* *Mantelliceras saxbii* *Neostlingoceras citanesis*	*Turkmenites gaurdakense*	*Turkmenites gaurdakense*	*Karamaites gaurdakensis Zone* *Mediasiceras beljakovae Bads*

Most of the species in this assemblage are endemic, except for *Acanthoceras rhotomagense asiaticum* Iljin and *A.* aff. *tapara* Iljin. *Kopetdagites kopetdagensis* was recorded in the lower part of the *Euomphalloceras cunningtoni* Zone of the middle Cenomanian (Atabekyan, 1986). This author also reported on the nominal subspecies of the *Acanthoceras* genus from the middle Cenomanian *Acanthoceras rhotomagense* Zone of the Kopetdag. *Acanthoceras rhotomagense* is an index species of the synonymous zone and occurs in England, France, Poland, Venezuela, Germany, Peru, and Romania. As mentioned, the ammonite assemblage includes a species *Acanthoceras* aff. *tapara* Wright somewhat differing from the type species. The holotype of this species was described from the middle Cenomanian deposits of northern Australia (Wright, 1963) and the Aitamir Formation (middle Cenomanian) of northern Iran (Seyed-Emami *et al.*, 1984). This testifies to the middle Cenomanian age of the *Kopetdagites aktaschensis* Zone. The latter can be correlated with the *Acanthoceras rhotomagense* and *Plesioacanthoceras amphibolum* zones of the western Kopetdag (Atabekyan, 1986) and also with the *Acanthoceras rhotomagense* Zone of the West European standard zonation (Kennedy, 1986).

Acanthoceras jukesbrownei Zone

This zone is characteristic of the the Gaurdak area of the Bukhara–Tajik region, being correlative to the middle part of the Gazdaganin Formation.

The ammonite assemblage consists of two species: *Acanthoceras jukesbrownei* Spath and *Newboldiceras newboldi* (Kossmat).

These species suggest the middle Cenomanian age of the zone. For instance, *Acanthoceras jukesbrownei* Spath was reported from the synonymous zone of England, France, Germany, Poland, and the former Soviet Union (Kopetdag), as well as from the middle Cenomanian of the United States. Species *Newboldiceras newboldi* (Kossmat) is a characteristic taxon of the *Acanthoceras jukesbrownei* Zone, the terminal one in the middle Cenomanian sequence of West Germany (Kaplan *et al.*, 1984). It also occurs in the middle–upper Cenomanian of England, France, Tunisia, Israel, China (Himalayas), Madagascar, and the interior United States. The species was also reported from the *Acanthoceras jukesbrownei* Zone of the western Kopetdag (Atabekyan, 1986). In southern India, the taxon in question is characteristic of the *Newboldiceras newboldi* Zone of the middle Cenomanian (Ayysami *et al.*, 1984). Based on these data, the *Acanthoceras jukesbrownei* Zone can be correlated with the analogous zone of West Europe (England, France, Germany) and Central Asia (western Kopetdag).

THE UPPER CENOMANIAN

Eucalycoceras pentagonum Zone

This zone is established in the Vakhsh and Southern Gissar areas of the Bukhara–Tajik region, being correlative to the lowermost upper Gazdaganin Formation.

The ammonite assemblage includes *Eucalycoceras pentagonum* (Jukes-Browne) and *Calycoceras* cf. *naviculare* (Mantell). Based on the stratigraphic range of these species, it can be correlated with the analogous zone of the western Kopetdag (Atabekyan, 1986) and, probably, with the *Calycoceras guerangeri* Zone of West Europe (Kennedy, 1986).

Sciponoceras gracile Zone

This zone is recognized in the Kafirnigan, Vakhsh, Southern Gissar, and Transalai areas of the Bukhara–Tajik region, as well as in the Fergana region (Alai, Naukat, Karadar'ya, and Kan areas), where it corresponds to the upper Gazdaginin Formation.

Its ammonite assemblage is represented by *Metoicoceras geslinianum* (Orb.), *Sciponoceras gracile* (Shumard), *Allocrioceras annulatum* (Shumard), *Anisoceras plicatile* (Sow.), *Euomphalloceras septemseratium* (Gragin), *Worthoceras vermiculum* (Shumard), *Vascoceras diartianum* (Orb.), *Metoicoceras nurekense* Iljin, *Sciponoceras baculoide* (Mantell), *Pseudocalycoceras* sp., *Placenticeras* cf. *memoriaschloenbachi* (Laube et Bruder), *Forbesiceras* ex gr. *obtectum* Sharpe, *Borissiakoceras orbicularum* Stephenson, and *B. mirabile* Arkh.

The above-listed species suggest the late Cenomanian age of the *Sciponoceras gracile* Zone. It can be correlated with the *Metoicoceras geslinianum* Zone of the standard ammonite scale of West Europe (Kennedy, 1986) and with the analogous zone of the interior United States (Gobban, 1984). It should be noted, however, that there is also an opinion that the *Sciponoceras gracile* Zone is of the early Turonian age. This viewpoint is shared by Atabekyan (*Postanovleniya...*, 1989).

THE TURONIAN STAGE

The lower Turonian

Watinoceras coloradoense Zone

This unit is identified in the Gaurdak, Kafirnigan, Vakhsh, Khozretishi, Obiniou, and Southern Gissar areas of the Bukhara–Tajik region, as well as in the Alai, Naukat, and Karadar'ya areas of the Fergana region. In all these areas, it is correlative to the lower part of the Talkhab Formation.

The zonal ammonite assemblage includes *Watinoceras coloradoense* (Henderson), *Puzosia chivensis* Arkh., *Proplacenticeras kharesmense* (Lahusen), and *Mammites nodosoides chivensis* Arkh. Iljin (1969) reported on the ammonite assemblage of *Watinoceras coloradoense* (Henderson), *W. reesidei* Warren, and

W. pressulum (Iljin) from the lower Turonian of the Srednyaya Amudar'ya area (Beshtyube Hills). These forms are accompanied by *Mammites nodosoides chivensis* Arkh. and *Metasigaloceras rusticum* Sow. The analysis of the geographic distribution of *Metasigaloceras rusticum* Sow. and *Mammites nodosoides* Schloth. reveals the following. Species *Metasigaloceras rusticum* Sow. is known from the lower Turonian of England (the *Mytiloides labiatus* Zone), as well as from the basal layers of the *Mammites nodosoides* Zone of France (High Normandy, Gard). *Mammites nodosoides* Schloth. is an index species of the *Mammites nodosoides* Zone in the lower Turonian stratotype. In addition, this species occurs in the lower Turonian deposits of Germany, the Czech Republic, Slovakia, France, Spain, Syria, Morocco, Tunisia, Madagascar, Peru, Colombia, Mexico, and northern areas of the interior United States (Colorado, Kansas). Geographic distribution of the listed ammonite species occurring together with *Watinoceras coloradoense* (Henderson) indicates that the discussed zone is of the early Turonian age. The stratigraphic position of this unit is similar to that in the lower Turonian stratotype section. In the latter, the *Watinoceras coloradoense* Zone corresponds to the base of the Turonian. The unit in question can be correlated with the analogous zone of southern England, France, the interior United States (Gobban, 1988), the Anglo-Parisian basin (Amedro, 1981), and West Europe (Kennedy, 1986). In some sections of the Gaurdak area, the endemic species *Proplacenticeras kharesmense* (Lahusen) coexists with *Mammites nodosoides chivensis* Arkh. and *Mytiloides labiatus* (Schloth.). The nominal subspecies *Mammites nodosoides nodosoides* represents an index taxon of the lower Turonian stratotype. Consequently, the early Turonian age of the zone is correct. Based on this evidence, the *Watinoceras coloradoense* Zone can be correlated with the synonymous zone of the interior United States and *Mammites nodosoides* Zone of the stratotype (the lower part).

Thomasites koulabicus Zone

Deposits of this zone correlative to the middle part of the Talkhab Formation conformably rest upon the *Watinoceras coloradoense* Zone. They are widespread in the eastern areas of the Bukhara–Tajik region (Vakhsh, Khozretishi, Obiniou, and Transalai areas) and in the Alai, Karadar'ya, and Naukat areas of the Fergana region.

The ammonite assemblage consists here of *Beschtubeites faasi* Arkh., *B.* cf. *alaiense* (Luppov), *B. kysylcurganense* (Luppov), *B. kutuzovae* Iljin, *B.* cf. *tenerum* Iljin, *Borissiakoceras* sp. indet., *Fagesia catinus* (Mantell), *Lewesiceras peramplum* Mant., *Puzosia* cf. *chivensis* Arkh., *Paramammites saenzi cassisianum* Thomel, *Nigeroceras kleri* Atabekjan, *Thomasites kuolabicus* (Kler), and *Proplacenticeras simakovi* (Luppov). The early Turonian age of the zone is substantiated by the occurrence of *Thomasites kuolabicus* (Kler), *Fagesia catinus* (Mantell), *Paramammites saenzi cassisianum* Thomel, and *Lewesiceras peramplum* (Mantell).

Forms of the *Thomasites* genus are common in many Tethyan regions: Algeria, Tunisia, Egypt, Jordan, Israel, Lebanon, Syria, Spain, southeastern France, Nigeria, and Texas (Reyment, 1954; Wright and Kennedy, 1981); the index species also occurs in the eastern regions of Central Asia and southeastern Nigeria (Zaborski, 1985). *Fagesia catinus* (Orb.) is characteristic of the lower Turonian stratotype. It is also widely distributed in the lower Turonian deposits of southern England, France, Venezuela, northern Mexico, the United States (Texas, Montana, California), Brazil, and Japan. The *Nigericeras* genus, first discovered in the former Soviet Union, is represented by the endemic species. However, the distribution of this genus shows that most of its species are characteristic of the lower Turonian of Sudan, Sahara, Niger, Algeria, Nigeria, Israel, the United States, England, and Angola. *Lewesiceras peramplum* (Mantell) is reported from the *Mammites nodosoides–Collignoniceras woolgari* Zone of England, France, Germany, the Czech Republic, and Slovakia. The assemblage also includes representatives of the endemic *Beshtubeites* genus, whose stratigraphic range is restricted to the lower Turonian.

Thus, the analysis of the above-listed species suggests the early Turonian age of the *Thomasites koulabicus* Zone. The latter can be correlated with the *Mammites nodosoides* Zone of the lower Turonian stratotype and the synonymous zone of England, the United States (Texas), Israel, Nigeria, and Spain (Cooper, 1979).

The middle Turonian
Spathites (Jeanrogericeras) reveliereanum Zone

This unit is recognized in the Gaurdak, Kafirnigan, Vakhsh, and Southern Gissar areas of the Bukhara–Tajik region, where it corresponds to the upper Talkhab and lower Dasgiryak formations.

Its ammonite assemblage is represented by *Spathites (Jeanrogericeras) reveliereanum* (Courtiller), *Neoptychites cephalotus* Courtiller, *Prohauericeras goupilianum* (Orb.), *Sciponoceras bohemicum* (Fritch), *Tragodesmoceras* sp., *Nurekites costatus* (Stankevich), *N. nurekensis* (Khakimov), *N. raricostatus* (Stankevich), *N. braecostatus* (Stankevich), *N. bulgariensis* Khakimov, *Lecointriceras fleurrisianum* (Orbigny), and *L. aksuensis* Iljin. The early Turonian age of the zone is inferred from the presence of the index species. The stratigraphic range of this species is restricted to the middle Turonian. In France (Touraine, Aquitaine, Provence), Spain, the Czech Republic, and Slovakia, it appears simultaneously with first specimens of *Collignoniceras woolgari* (Mant.) and *Kamerunoceras turoniense* (Orb.). In the Turonian stratotype (Touraine), this species occurs in the *Kameruno-*

ceras turoniense Zone, where its maximum abundance is confined to the upper part of the latter. The taxon in question was also reported from the lower part of the middle Turonian *Pseudospidoceras subconciliatum* Zone of the southern Pyrenees (Bilote, 1981).

The analysis of the geographic and stratigraphic distribution of the above-mentioned species indicates the middle Turonian age of the *Spathites (Jeanrogericeras) reveliereanum* Zone. The unit is correlative to the middle Turonian *Kamerunoceras turoniense* Zone (Robaszynski, 1983), the synonymous zone of the Turonian stratotype, and the *Pseudospidoceras subconciliatum* Zone of the southern Pyrenees (Bilote, 1981).

Collignoniceras woolgari Zone

This zone correlative to the upper Dasgiryak Formations is identified in all the areas of the Bukhara–Tajik region. It is also recognized in the Pamirs (the Rangkul' area) and in the Alai, Karadar'ya, and Naukat areas of the Fergana region. In the Magian area of the Zeravshan–Gissar region, this unit corresponds to lower beds of the Gezan Formation.

The ammonite assemblage consists of *Collignoniceras woolgari* (Mantell), *C. bakeri* (Anderson), *C. carolinum* (Orb.), *C. turoniense* (Sornay), *C. canthus* (Sornay), *Lecointriceras fleurrisianum* (Orb.), *L. iljini* Khakimov, *L.* cf. *costatum* Kennedy, Wright et Hancock, *Pseudoforresteria charrochensis* Khakimov, *P. babatagensis* Khakimov, *Prionocyclus* cf. *hyatti* (Stanton), and *Romaniceras* cf. *kallesi* (Zazvorka).

The ammonite assemblage suggests the middle Turonian age of the *Collignoniceras woolgari* Zone. The lower and middle parts of the zone can be correlated with the upper part of the *Kamerunoceras turoniense* Zone (Robaszynski, 1983) known in England and Turonian type areas of France (Kennedy *et al.*, 1981). The presence of *Romaniceras kallesi* and *Prionocyclus hyatti* in the upper layers of the *Collignoniceras woolgari* Zone indicates that these beds likely correspond to the *Romaniceras kallesi* Zone of West Europe (Amedro and Hancock, 1985). This part of the zone can probably be correlated also with the lower portion of the *Prionocyclus hyatti* Zone of the western interior United States (Cobban, 1984), because it encloses the zonal species *Romaniceras kallesi* (Zazvorka). The latter provides grounds to correlate the lower part of the *Prionocyclus hyatti* Zone with the *Romaniceras kallesi* Zone of West Europe.

The upper Turonian Subprionocyclus neptuni Zone

This unit is distinguished in the Gaurdak and Kafirnigan areas of the Bukhara–Tajik region, where it corresponds to the Muzrabat Formation.

The ammonite assemblage includes *Pseudoforresteria ornata* Iljin, *P. asiatica* Khakimov, *Prionocyclus cobbani* Matsumoto, *P.* cf. *wyomingensis* Meek, *Proplacenticeras* cf. *arkhangelski* Iljin, *Subprionocyclus branneri* (Matsumoto), *S. beschtubense* Iljin, *S.* cf. *normalis* (Anderson), *S. neptuni* (Geinitz), *S. inflatus* Khakimov, *S. pitniakensis* Iljin, *S. bravaisianus* (Orbigny), *S. gissarensis* Khakimov, and *Lewesiceras sharpei* Spath.

Despite its provincial composition, this assemblage substantiates the late Turonian age of the *Subprionocyclus neptuni* Zone. Based on the presence of index species, *S. branneri*, and *S. normalis*, it is correlated with the analogous zone of West Europe, whereas the occurrence of *Prionocyclus* cf. *wyomingensis* allows this unit to be correlated with the *Prionocyclus wyomingensis* Zone of the interior United States (Cobban, 1984).

Coilopoceras gissarensis Beds

This unit is identified in the Gaurdak and Vakhsh areas of the Bukhara–Tajik region and corresponds to the lower Modun Formation.

The ammonite assemblage is represented by two species: *Coilopoceras gissarensis* Iljin and *Placenticeras akrabatense* Vinokurova. The *Coilopoceras* genus is represented by the endemic species. However, most of the known species of this genus (approximately 80–90%) are the Turonian in age. Its representatives occur in the upper Turonian deposits of many regions of the world: the United States (California), Mexico, Trinidad, Venezuela, Colombia, Peru, Ecuador, Brazil, France, Lebanon, Algeria, Nigeria, Angola, Madagascar, and India. The stratigraphic position of *Coilopoceras gissarensis* Iljin is similar to that of other representatives of the *Coilopoceras* genus in West European sections. *C. inflatum, C. colleti,* and *C. springeri* from the interior United States occur at a stratigraphic level (in *Prionocyclus hyatti* and *Prionocyclys macombi* zones) lower than that of *Coilopoceras gissarensis*.

Hourcquia pacifica Zone

This zone corresponds to the upper Modun Formation in the Gaurdak, Kafirnigan, and southern Gissar areas of the Bukhara–Tajik region.

Its ammonite assemblage includes *Hourcquia pacifica* Matsumoto, *H.* aff. *hataii* Hashimoto, *H.* cf. *mirabilis* Collignon, *Aktaschia akrabatense* (Iljin), *Proplacenticeras kotzi* Iljin, *P. proplanum* Iljin, *P.* cf. *orbignyanum* (Geinitz), *Prionocyclus* cf. *quadratus* Cobban, and *Reesidites minimus* (Hausaka et Fukada).

The known stratigraphic ranges of these species suggest the late Turonian age of deposits in question and allows them to be correlated with the *Hourcquia pacifica, Subprionocyclus normalis*, and *Reesidites minimus* zones of Japan (Matsumoto and Noda, 1966; Matsumoto, 1981; Kennedy, 1986).

THE CONIACIAN

Prionocycloceras guayabanum Zone

The *Prionocycloceras guayabanum* Zone is recognized in the Gaurdak and Vakhsh areas of the Bukhara–Tajik region, where it corresponds to the lower and middle parts of the Akrabat Formation.

The ammonite assemblage comprises *"Lewesiceras" asiaticum* Iljin, *Prionocycloceras guayabanum* (Stein.), *P.* cf. *multicostatum* Collignon, *P.* aff. *multicostatum* Collignon, *Proplacenticeras orbignyanum* (Geinitz), *P. proplanum* Iljin, *Placenticeras pitniakense* Iljin, *Peroniceras* aff. *proteus* (Matsumoto), and *Prionocycloceras* aff. *carvalhoi* (Howarth). The stratigraphic ranges of these species indicate the Coniacian age of the zone. *Inoceramus percostatus* Müller recorded in the Coniacian deposits of the Aktash section in the Gaurdak area is of great importance for correlation. This species occurs in the upper Coniacian *Gauthiercras margae* Zone of northern Germany, the upper layers of the upper Coniacian *Inoceramus involutus* Zone of the Northern Caucasus, upper horizons of the lower Coniacian *Inoceramus wandereri* Zone of the Kopetdag, and the upper Coniacian *Inoceramus involutus* Zone of the western Kopetdag (Atabekyan, 1966). The equivalent status of stratigraphic ranges of all these units requires further confirmation. By the presence of index species, the *Prionocycloceras guayabanum* Zone can be correlated with the *Prionocycloceras guayabanum* and *Gauthiericeras margae* zones of Madagascar, whereas the occurrence of *Peroniceras* and *Forresteria* forms in the first unit allows it to be correlated with standard *Forresteria* (*Harleites*) *petrocoriensis* and *Peroniceras* (*Peroniceras*) *tridorsatum* ammonite zones of West Europe (Kennedy, 1984).

Eostantonoceras babatagensis Beds

This unit corresponds to the upper Akrabat Formation and is identified in the Gaurdak, Kafirnigan, and Vakhsh areas of the Bukhara–Tajik region.

Ammonites are represented only by two provincial species: *Eostantonoceras babatagensis* Khakimov and *E. crassum* (Iljin). The Coniacian age of deposits in question is inferred from their stratigraphic position. They are underlain by deposits of the *Prionocycloceras guayabanum* Zone and overlain by the Santonian layers, whose age is inferred from the occurrence of representatives of the *Stantonoceras* genus. More evidence favoring their Coniacian age is the presence of forms belonging to the new *Eostantonoceras* genus occupying an intermediate position between the *Stantonoceras* and *Placenticeras* genera. The unit can arbitrarily be correlated with the upper Coniacian *Gauthiericeras margae* and *Paratexanites serratomarginatus* zones (Kennedy, 1984).

THE SANTONIAN

The lower Santonian
Local Stantonoceras guadalupae asiaticum Zone

This unit, correlative to the lower and middle parts of the Kattakamysh Formation, is characteristic of both the Bukhara–Tajik and the Fergana regions.

Its ammonite assemblage consists of *Placenticeras bobkovae* Iljin, *P. luppovi* Iljin, *Stantonoceras guadalupae asiaticum*, and *S. kysylcumense* Arkh.

Iljin (1969) reported that *Stantonoceras kysylcumense* Arkh. occurs together with *Inoceramus patchi* Arkh. in the Koshabulak section of the Srednyaya Amudar'ya area. Forms of this genus are characteristic of the lower Santonian deposits of the East European platform and the *Inoceramus undulatus* and *Inoceramus cordiformis* zones of Poland, Germany, and western areas of Central Asia, except the Mangyshlak, Ustyurt, and southern Aral regions (Atabekyan, 1986). They also occur in the *Inoceramus undulatoplicatus* and *Inoceramus cordiformis* zones of Germany. In the phylogenetic respect and according to the evolutionary level of suture lines, the subspecies *Stantonoceras guadalupae asiaticum* is a predecessor of the nominal subspecies *S. guadalupae guadalupae*, which occurs in the uppermost Santonian–lower Campanian layers of North America (Iljin, 1969).

Thus, it appears that the local *Stantonoceras guadalupae asiaticum* Zone should be correlated with the *Texanites quinquenodosus* and *Hemitissotia randoi* zones of the lowermost Santonian in Austria (Gosau) (Summesberger, 1983).

Stantonoceras polyopsis Zone

The *Stantonoceras polyopsis* Zone that corresponds to the middle Kattakamysh Formation is recognized in the Gaurdak area of the Bukhara–Tajik region and Srednyaya Amudar'ya area of the interior Central Asia.

Only one ammonite species *Stantonoceras polyopsis* occurs in deposits of this zone. The species is widespread in the Santonian deposits of Austria. It appears in the M^2 and N^2 beds of Arnou's scheme for the Aquitaine and also occurs in the *Marsupites testudinarius* Zone of northwestern Europe. This species is also recorded in France (Aquitaine, Provence, Corbiere regions), Germany, and the United States (Kennedy and Wright, 1983). Based on the occurrence of the index species, the zone can be correlated with the analogous zone of Austria (Summesberger, 1983).

The upper Santonian
Local Asiatostantonoceras tagamense Zone

The local *Asiatostantonoceras tagamense* Zone is recognized in the Gaurdak, Kafirnigan, and Southern Gissar areas of the Bukhara–Tajik region and interior

areas of Central Asia, where it correspods to the upper part of the Kattakamysh Formation.

This unit is distinguished by occurrence of ammonite species *Asiatostantonoceras tagamense*. According to Iljin's data (1969), this species is close to ammonite forms from the *Marsupites testudinarium* Zone of Germany described by Schlöter as *Ammonites syrtalis* var. *polyopsis* Schlöter (non Dujardin) and *Ammonites syrtalis* var. *orbignyanus* Schlöter (non Geinitz). Moreover, species of the *Asiatostantonoceras* genus are characteristic of upper layers of the Santonian in Germany (Müller and Wolleman, 1906). More evidence favoring the late Santonian age of the *Asiatostantonoceras tagamense* Zone is the fact that it is conformably underlain and overlain by the ammonite-substantiated lower Santonian and lower Campanian deposits, respectively.

THE CAMPANIAN

The lower Campanian
Scaphites inflatus Zone

This zone, which corresponds to the lower part of the Sarakamysh Formation, is established only in the Gaurdak area of the Bukhara–Tajik region.

The ammonite assemblage is represented here by *Scaphites inflatus* Roem., *Gissarites tagamense* Iljin, and *G. kysylchense* Iljin. The early Campanian age of the unit is inferred from the presence of the index species. The latter is characteristic of the *Scaphites binodosus* Beds in western Germany coeval with the lower Campanian *Diplomoceras bidorsatum* and *Menabites campaniense* zones in the stratotype section (Iljin, 1969).

The upper Campanian
Hoplitoplacenticeras marroti Zone

This unit is distinguished in the Gaurdak, Kafirnigan, and Vakhsh areas of the Bukhara–Tajik region and in the Kan area of the Fergana region; it is correlative to the lower Daralitau Formation.

Its ammonite assemblage includes *Hoplitoplacenticeras (Hoplitoplacenticeras) bucharense* Iljin, *H. (H.) marroti* (Goq.), *H. (H.) vari* (Schlöter), *Hoplitoplacenticeras (Lemfordoceras) rarecostatum* Khakimov, *Trachyscaphites spiniger* (Schlüter), and *T. gibbus* (Schlüter).

The stratigraphic and geographic ranges of the species suggests the late Campanian age of the zone. Based on the presence of index and other species, it can be correlated with the *stobaei–basian* and *conica–mucronata* subzones of Braunschweig–Hannover (Ernst *et al.*, 1979), the *Trachyscaphites spiniger* Zone of northwestern Germany (Wiedman, 1979), *Bostrychoceras polyplocum* of Germany, the *Neancycloceras phaleratum* and *Bostrychoceras polyplocum* zones of Poland (Blaszkiewicz, 1980), and *Belemnitella langei* Zone of the East European platform (Naidin, 1979), as well as with the *Hoplitoplacenticeras ceesfeldiense–Stegaster gillieroni* Zone and the lower part of the *Bostrychoceras polyplocum* Zone of the Kopetdag (Atabekyan, 1979).

Trachyscaphites pulcherrimus Zone

This unit corresponds to the middle Daralitau Formation and is characteristic of the Kafirnigan, Vakhsh, and southern Gissar areas of the Bukhara–Tajik region.

The ammonite assemblage consists of *Didymoceras vancouverensis* Gabb., *D. parastevensoni* Atabekjan et Khakimov, *Trachyscaphites spiniger spiniger* (Schlöter), *T. pulcherrimus* (Roemer) *Glyptoxoceras retrosum* Schlüter, *G. mortoni* Meek et Hauden, *Baculites anceps* Lam., and *B. vertebralis* Lam.

The index species is present in the *Bosrtychoceras polyplocum* Zone of northwestern Germany (Wiedman, 1979), the *Bosrtychoceras polyplocum* and *Didymoceras donezianum* zones of Poland representing the middle part of the upper Campanian (Blaszkiewicz, 1980), the uppermost Campanian *Belemnitella langei* Zone of the East European platform (Naidin, 1979), and in the upper Campanian beds of central and eastern Kopetdag.

It was also reported from the upper Campanian–Maastrichtian (the Navarro Formation) of the United States (Arkansas, Colorado, Tennessee, California, Texas; Cobban, 1974) and in the upper Campanian of Israel (Lewy, 1969) and Colombia (Petters, 1976). The provincial species *Didymoceras parastevensoni* Atabekjan et Khakimov was identified from the *Bosrtychoceras polyplocum* Zone in the western Kopetdag (Atabekyan, 1986).

Thus, the ammonite assemblage in question indicates the late Campanian age of the *Trachyscaphites pulcherrimus* Zone and allows it to be correlated with the *Didymoceras donezianum* Zone of Poland (Blaszkiewicz, 1980), the *Belemnitella langei* Zone of the East European platform (Naidin, 1979), and the *Hoplitoplacenticeras vari–Micraster glyphus* Zone of central and eastern Kopetdag.

THE MAASTRICHTIAN

The lower Maastrichtian
Nostoceras hyatti Zone

This zone is recognized in the Kafirnigan area of the Bukhara–Tajik region, where it corresponds to the upper Daralitau Formation.

The ammonite assemblage includes *Nostoceras hyatti* (Stephenson), *N.* cf. *draconis* (Stephenson), *N.* cf. *navarroensis* Shumard, *N.* cf. *pozaryskii* Blaszkiewicz, *Neohamites subcompressus* (Forbes, *Ieletskytes* aff. *pumilus* (Stephenson), *I. brevis* (Meek), *I.* aff. *brevis* (Meek), and *Baculites anceps* Lam.

Until recent time, the considered zone was referred to the late Campanian (Khakimov, 1990). However, the analysis of the stratigraphic range of ammonite species from this zone suggests the early Maastrichtian age of

the latter. For instance, the index species was reported from the lower Maastrichtian of the Aquitaine basin, where it is characteristic of the R_2 Horizon (Maastrichtian) in the Arnau' zonation. In addition, this species was recorded in the lower Maastrichtian of the interior United States and Angola (Cobban, 1974), as well as in the uppermost Campanian–lower Maastrichtian of Israel (Lewy, 1969). A juvenile specimen of the *Nostoceras* genus was found in the *Caster* Horizon (upper Maastrichtian) of the stratotype section (Kennedy, 1986). *Didymoceras schloenbachi* (Favre) was described from the lower Maastrichtian of Poland and Ukraine (the Lviv depression) and from the Maastrichtian of Bulgaria (Kennedy and Sumesberger, 1987). This species was also identified, along with *Didymoceras navarroensis* Shumard, in the *Hauericeras sulcatum* Zone (lower Maastrichtian) of the western Kopetdag (Atabekyan, 1986). Its stratigraphic range spans the *Belemnella lanceolata, Belemnella pseudoobtusa*, and *Belemnella obtusa* belemnite zones of the lower Maastrichtian (Blaszkiewicz, 1980). *Glyptoxoceras subcompressum* (Forbes) is characteristic of Cunred Limestone of the upper Maastrichtian of Limburg (Kennedy, 1986).

Thus, the ammonite assemblage in question testifies to the early Maastrichtian age of the *Nostoceras hyatti* Zone. This is also supported by the radiolarian assemblage whose renovation coincides with the lower boundary of the zone: most radiolarian species here show the Maastrichtian affinity.

By the presence of common species (*Didymoceras schloenbachi* (Favre) and *D.* cf. *navarroensis* Shum.), the unit corresponds to the *Hauericeras sulcatum* Zone in the western Kopetdag as well as to the *Belemnella lanceolata, Belemnella pseudoobtusa,* and *Belemnella obtusa* belemnite zones of the lower Maastrichtian (Schulz and Schmid, 1983).

Baculites knorrianus Zone

This zone corresponds to the Udantau Formation and is established in the Kafirnigan, Khozretishi, and Obiniou areas of the Bukhara–Tajik region.

Only two ammonite species characterize this unit: *Baculites anceps* Lam. and *B. knorrianus* Desmarest. *Baculites anceps* Lam. was reported from the Maastrichtian deposits of Italy, Chile, and Argentina, as well as from the lower Maastrichtian of France (Normandy), Sweden, Holland (Limburg), Germany, England, Poland, and Japan. In Angola, this species was found in the upper Campanian; in the territory of the former Soviet Union, its range covers the upper Campanian–lower Maastrichtian. The species was also recorded in the Maastrichtian of West Africa, Belgium, Germany, Denmark, Sweden, Bulgaria, Poland, the Lviv region, Crimea, Caucasus, Transcaspian region, and West Siberia (Naidin, 1974). Kennedy (1986) reported this species to be present in the upper Maastrichtian of Denmark, Holland (Cunred, Limburg, Gulchem, Maastricht), the former Soviet Union, and, probably, Spain. *Baculites knorrianus* Desmarest was identified in the lower Maastrichtian of the former Soviet Union (Naidin, 1974), northern Poland, Germany, France, England, Sweden, Belgium, Holland, Spain, the United States, and northwestern Bulgaria (Naidin, 1974; Kennedy and Summesberger, 1987). Birkelund (1957) noted this species in the lower Maastrichtian and basal layers of the upper Maastrichtian of Holland. I first discovered *Baculites knorrianus* in the western Aktash section of the Kafirnigan area, where it occurs together with larger foraminifers *Siderolites calcitropoides* Lam. and *Orbitolites media* Arch (Dzhalilov *et al.*, 1988). Eastward, coeval deposits host a similar foraminifer assemblage but are barren of ammonite fossils (Gol'tman and Ashurov, 1989). Based on larger foraminifers, the *Baculites knorrianus* Zone is traced farther to the eastern areas of Central Asia.

Thus, the ammonite assemblage in question and larger foraminifers indicate the Maastrichtian age of the unit.

The occurrence of the index species allows the *Baculites knorrianus* Zone to be correlated with the upper part of the Maastrichtian *Belemnella occidentalis* Zone (Birkelund, 1957).

CONCLUSION

(1) Based on ammonite succession, a stratigraphic scale was proposed for the Upper Cretaceous of Central Asia. New stratigraphic units introduced allow it to be correlated with the standard stages of stratotypes;

(2) It was established that ammonoids occur in all the stages of the Upper Cretaceous. Their maximum diversity is characteristic of the lower Turonian (18 species) and upper Campanian (17 species);

(3) Cenomanian ammonoids from eastern Central Asia reveal the greatest taxonomic similarity with ammonoids of West Europe, Tunisia, Israel, Iran, and China (Himalayas);

(4) Based on the study of ammonite assemblages, three-member subdivision of the Cenomanian and Turonian was first substantiated. It was established that the upper boundary of the Turonian Stage is substantially younger than was accepted earlier;

(5) The lower boundary of the Maastrichtian is placed at a lower level in comparison with previous scales;

(6) Twenty-four stratigraphic units were identified instead of the thirteen units in the previous zonation.

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Substantiation of the Paleocene–Eocene Boundary in Western Siberia by Foraminifers

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Abstract—Data on stratigraphy and foraminifers from the Paleocene and Eocene deposits in western Siberia are correlated and the boundaries between the two series are verified. New results for boreholes 1, 2, 3, and 4 on the Kaimysov arch (the Central region) are presented. Judging from the microfauna composition, especially from that of endemic agglutinated siliceous foraminifers and from peculiar features of sedimentation, the West Siberian province is assumed to represent the southern margin of the biogeographic Arctic Circumpolar Realm of the discussed epochs. The benthonic calcareous and planktonic foraminifers occasionally encountered in the province were correlated with similar Paleocene and Eocene forms, among them those from stratotypes of Denmark, Belgium, and England. The Danian (the *Brotzenella praeacuta* Zone of the Lower Paleocene) of western Siberia is established in the upper layers of the Gan'kino Formation containing benthonic and planktonic foraminifers. The lower half of the Talitza Formation, within which the *Ammoscalaria friabilis* Zone has been distinguished, is referred to the Zealandian. In the Transurals region, this part of the formation corresponds to the *Cibicidoides proprius* Beds, many species of which are encountered in appropriate deposits of the Zealand Island stratotype and southern Sweden. The Thanetian *Cibicidoides favorabilis* Zone is recognized on the basis of planktonic foraminifers found in the Turgai depression, along with its zonal species, and the unit is correlated with that of the Thanet Island stratotype (England). The Ypresian (Lower Eocene) is established in western Siberia on the basis of calcareous benthonic and planktonic foraminifers. The *Anomalinoides ypresiensis* Zone, similar to the synonymous zone of the Ypresian in the Netherlands, is distinguished for the first time in this part of the section. The zone is confidently defined in sections of boreholes 1 and 2 of the Kaimysov arch. In addition to benthonic forms, *Planorhotalites* species typical of the Ypresian are found. The boundary between the Paleocene and Eocene is distinguished between the Thanetian *Cibicidoides favorabilis* and Ypresian *Anomalinoides ypresiensis* zones of western Siberia. The figure presented shows the correlation of the Paleocene and Eocene zones (boreholes 1, 2, 3, and 4) within the Kaimysov arch. Two micropaleontological plates are presented as well.

Key words: *western Siberia, stratigraphy, correlation, boundaries, Paleocene, Eocene, stages, microfaunal zones.*

Data on foraminifers widespread in western Siberia are basic in the study. Radiolarians encountered in abundance together with foraminifers were also studied. The problem of the Paleocene and Eocene boundary in the region depends on subdivision of the Paleocene into stages, which is a complicated challenge because of the endemic character of many taxons of the West Siberian microfauna.

Foraminifers widespread in different paleocoenoses are represented mainly by agglutinated, more rarely calcareous, benthonic and planktonic forms. Local zones and beds, originally distinguished on the basis of this fauna, were dated to the Danian–Paleocene time (Dain, 1961; Kiselman, 1960, 1978; Lipman *et al.*, 1960; Podobina, 1975, 1984; Subbotina and Kiselman, 1961; Subbotina *et al.*, 1964; Freiman, 1960; and others). Recent data on foraminifers, along with generalized data on stratigraphy, paleobiogeography and paleogeography, as well as the revealed correlation levels corresponding to world-wide eustatic cycles, made possible great progress in the solution of the presented problem.

By the microfauna composition and sedimentation patterns, I referred the West Siberian province to the southern margin of the paleobiogeographic Arctic Realm of the Paleocene–Eocene time. However, in the Transurals region and in the southwest of the province, there are isolated calcareous benthonic and planktonic foraminifers, which are similar to some West European species, including those from stratotypes of Denmark, Belgium, and England.

The sedimentation of the Danian time is closely related to the cessation of the Cretaceous carbonate sedimentogenesis and represents the final stage of the carbonate cyclothem (Naidin and Ben'yamovskii, 1989). The nature of the fauna and, especially, the presence of Paleogene planktonic foraminifers provided the basis for assigning this stage to the Paleogene. The

Danian deposits (the *Brotzenella praeacuta* Zone) are distinguished in the uppermost layers of the Gan'kino Formation of Western Siberia. The deposits survived erosion mainly in depressions (Omsk, Ust-Tyma, and other localities).

In addition to the abundant Maastrichtian species, characteristic forms of the *Brotzenella praeacuta* Zone include *Gaudryina gigantica* (Subbotina), *Clavulina parisiensis* Orb., *Lenticulina inusitata* Kisselman, *Parrella lens* (Brotzen), *Cibicides spiropunctatus* Galloway et Morrey, *Anomalinoides danicus* (Brotzen), and *Brotzenella praeacuta* (Vassilenko).

Besides the above-mentioned benthonic foraminifers, the zone comprises rare planktonic foraminifers *Globigerina varianta* Subbotina, *G. trivialis* Subbotina, *G. pseudobulloides* Plummer, and others. Secretory calcareous benthonic forms with dominant rotaliids prevail throughout the zone. In the eastern areas of West Siberia (the Ust-Tyma depression), this part of the section correlates with beds whose paleocenosis includes *Bathysiphon nodosarieformis* and *Glomospira charoides* (Podobina, 1992).

The Zealandian corresponds to the onset of a new eustatic transgression and the formation of the overlying terrigenous–siliceous cyclothem. The transgression is recorded in the "Copenhagen Paleocene" deposits (the stratotype of the Zealandian in Denmark, Zealand Island), the Paleocene strata of southern Sweden (Malmö Mount), the lower Syzran deposits of the Volga region, and the Talitsa Formation of western Siberia. Zones of planktonic foraminifers corresponding to the Paleocene stages represent chronozones of the general stratigraphic scale (Krasheninnikov and Basov, 1986; *Reshenie* ..., 1989). The taxonomic composition of planktonic and benthonic foraminiferal assemblages in these zones represents a change in the paleogeographic environment of sea basins related to the Cenozoic sea-level fluctuations and other factors. Many researchers interpret a sudden change in the Paleocene eustatic curve at the boundary of two cyclothems as separating zones of *Acarinina inconstans* (the uppermost Danian) and *Morozovella angulata* (the lowermost Zealandian). Great changes in the taxonomic composition of planktonic foraminifers at this datum level led Krasheninnikov (1982), Harland (1985), and other researchers to place the boundary between the lower and upper Paleocene here. In the above-mentioned works, the *Morozovella angulata* and *M. conicotruncata* zones correspond to the Zealandian rather than to the Montian (*Reshenie*..., 1989). This should be taken into account during further stratigraphic studies of this part of the section, as decided at the 28th International Geological Congress, Washington, 1989.

Terrigenous–siliceous deposits of the Talitsa Formation are widespread western Siberia, and, as noted above, they correspond to the Zealandian transgression. The formation comprises abundant siliceous foraminifers of the paleocenosis with *Ammoscalaria friabilis* of the relatively deep-water type. The dominant species here are *Labrospira granulosa* (Lipman), *Haplophragmoides periferoexcavatus* Subbotina, *Cyclammina coksurovae* Uschakova, *Ammoscalaria friabilis* (Ehremeeva), *Verneuilinoides paleogenicus* (Lipman), and *Trochammina pentacamerata* Lipman. The Talitsa Formation, which comprises the *Ammoscalaria friabilis* assemblage, is recognized as a synonymous zone. Interlayers with calcareous forms of the shallow-water *Cibicidoides proprius* paleocoenosis are found in the Talitsa Formation sequence among layers with agglutinated forms. In this paleocoenosis, the following species are determined: *Bolivinopsis scanica* Brotzen, *Robulus discus* (Brotzen), *Eponides lunatus* Brotzen, *Gavelinella lellingensis* Brotzen, *Cibicidoides succedensis* Brotzen, *C. proprius* Brotzen, *Planulina limbata* Brotzen, *Nonionellina ovata* Brotzen, and *Bulimina ovata* Orb. In the southeastern part of West Siberia (the Ust-Tyma depression, boreholes 1, 2, 3, 152), the Zealandian *Ammoscalaria friabilis* Zone is correlated with the *Cyclammina coksuvorovae* Beds (Podobina, 1990; see Fig. 1). Foraminifers of this paleocoenosis are similar in species composition to the assemblage of the "Copenhagen Paleocene" (the stratotype of the Zealandian), as well as to the Paleocene assemblage of southern Sweden (Brotzen, 1948). I managed to correlate the assemblages thanks to samples submitted by Naidin and European researchers. The West Siberian assemblage with *Cibicidoides proprius* is impoverished as compared to the those from northern Europe and probably reflects a lower temperature of its basin, because of the influence of the Arctic region, and different conditions of sedimentation.

The Thanetian deposits in western Siberia are revealed supposedly in the upper layers of the Talitsa Formation and lower layers of the Lyulin-Vor Formation. The *Cibicidoides favorabilis* Zone is distinguished in this part of the sequence; the paleocoenosis of the zone comprises agglutinated siliceous forms of the *Reophax, Labrospira, Haplophragmoides,* and *Cyclammina* genera, which are abundant in the underlying zone. In addition to them, there appear singular new species of secretory calcareous forms, the zonal species included. In the southeast, the *Glomospira gordialiformis* and *Cyclammina coksuvorovae* beds correspond to the Thanetian *Cibicidoides favorabilis* Zone (Lipman *et al.,* 1960; Podobina, 1990; Subbotina *et al.,* 1964). Amon and I (1992) revealed analogues of this zone in the Sarbai section of the Turgai depression (near the town of Rudnyi) and found calcareous benthonic and isolated planktonic forms of the Thanetian *Acarinina subsphaerica* Zone. Naidin and Ben'yamovskii (1989) established the *Acarinina subsphaerica–Cibicides reinholdi* Zone at this stratigraphic level in a series of sections of the Turgai depression. I have studied the Thanetian foraminifer assemblage (the Thanet Island locality) sent to me from England by Dr.D.Carry and referred it to the upper Paleocene. In the stratotype assemblage, I identified species of planktonic foramin-

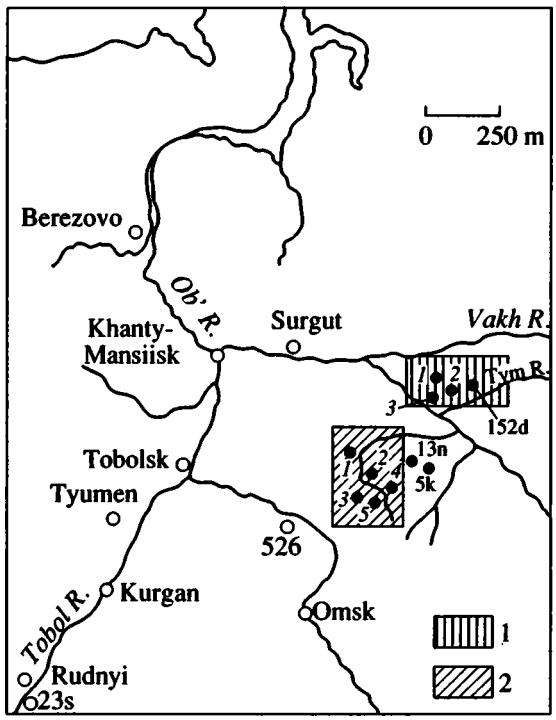

Fig. 1. Borehole localities: (1) boreholes *1, 2, 3,* and 152d drilled in the basins of the Tyma and Paidugina rivers (Paidugina Group of the Tomsk Geological and Exploration Expedition); (2) boreholes *1–5* drilled in the basin of the north–south trending course of the Vasyugan River (Zapadnaya Group of the Tomsk Geological and Exploration Expedition); borehole (13n) drilled in the Nyurolka River basin (the Nyurolka Group of the Tomsk Geological and Exploration Expedition); borehole (5k) drilled in the Chizhapka River basin (Chizhapka Group of the Tomsk Geological and Exploration Expedition); borehole (526) drilled in the vicinity of the village of Zaviyalovo.

ifers *Globigerina incisa* Hillebrandt, *Acarinina intermedia* Subbotina, *A. soldadoensis* (Bronnimann), *A. trichotrocha* (Loeblich et Tappan), and others.

The analysis of the taxonomic composition of the planktonic foraminifers (the Thanetian stratotype, Thanet Is., England) shows that, according to the scale of 1989 of the International Stratigraphic Committee, they correspond to the *Acarinina subsphaerica* and *A. acarinata* zones. The Paleocene transgression reached England supposedly at the end of this epoch. Thus, there is no point in ranging the Thanetian through the whole upper Paleocene, as some researches did (Harland *et al.*, 1985).

The Paleocene marine deposits are widespread in Europe and Asia. Within western Siberia, specimens of planktonic and secretory benthonic species are rare in the Eocene deposits (Plates I, II), whereas agglutinated siliceous foraminifers are more diverse and abundant. For comparison, I studied the Eocene foraminiferal coenoses from the Turgai depression and found that their secretory calcareous species coexisted with agglutinated siliceous forms similar to those from West Siberia.

The Ypresian (lower Eocene) deposits with planktonic and benthonic calcareous foraminifers were originally distinguished by Aizenshtat (1964) in the Turgai depression as the *Acarinina pentacamerata* Zone.

Collections of Paleogene foraminifers sent to me by Ben'yamovskii, Zinovieva, and Faizulina made it possible to study benthonic secretory calcareous forms from this zone. I described the assemblage of benthonic foraminifers from the Belinskii section of the Turgai depression as the *Anomalinoides ypresiensis–Cibicides carinatus* paleocoenosis. It is similar to the *Anomalina (Anomalinoides) ypresiensis–Gaudryina hilltermanni* assemblage distinguished in the lower Eocene deposits of the Netherlands (Doppert and Neele, 1983). Some benthonic species from this stratigraphic level of the Turgai depression are known in the London Clay of England (Murrey and Whight, 1974). Among these, there are *Cibicides cunobelini* Haynes, *C. mauricensis* Howe et Roberts, *C. westi* Howe, and *Uvigerina batjesi* Kaasschieter. In West Siberia, this stratigraphic datum is correlated with the upper layers of the lower Lyulin-Vor Subformation with rare foraminifers and radiolarians.

Benthonic and planktonic foraminifers from the upper layers of the lower Lyulin-Vor Subformation, which have been detected in borehole sections of the central region (the Kaimysov arch, boreholes 1, 2, 3, and 4; see Fig. 2), are of interest. I distinguished the layers bearing the benthonic calcareous and agglutinated siliceous forms together with the characteristic *Anomalinoides ypresiensis* species as the synonymous zone of the Ypresian (Lower Eocene).

The borehole section 1 (Zapadnaya Group, the Kaimysov arch) revealed the most complete characteristics of the microfauna. In addition to rare agglutinated

Plate I. Foraminifers from the lower Eocene deposits of West Siberia: (a) back or side view; (b) ventral view; (c) aperture view; magnification ×100 for all specimens.

(1) *Anomalinoides ypresiensis* (ten Dam), sample no. 1530, BH 1, depth 450 m, Ypresian; (2) *Acarinina pentacamerata* Subbotina, sample no. 1532, BH 1, depth 450 m, Ypresian; (3, 4) *Planorotalites pseudoscitulus* (Glaessner): (3) sample no. 1533, (4) sample no. 1534. BH 1, depth 470 m, Ypresian.

All specimens are stored in the collection of the Micropaleontology Department, Paleontological Museum, Tomsk State University. The specimens are from the lower subformation of the Lyulin-Vor Formation (the synonymous horizon of the Lower Eocene) recovered by Borehole 1 in the middle part of the north–south trending course of the Vasyugan River (Zapadnaya Group, the westernmost area of Tomsk oblast), West Siberia.

Drawings from natural specimens were made by artist O.M. Lozovaya.

Plate I

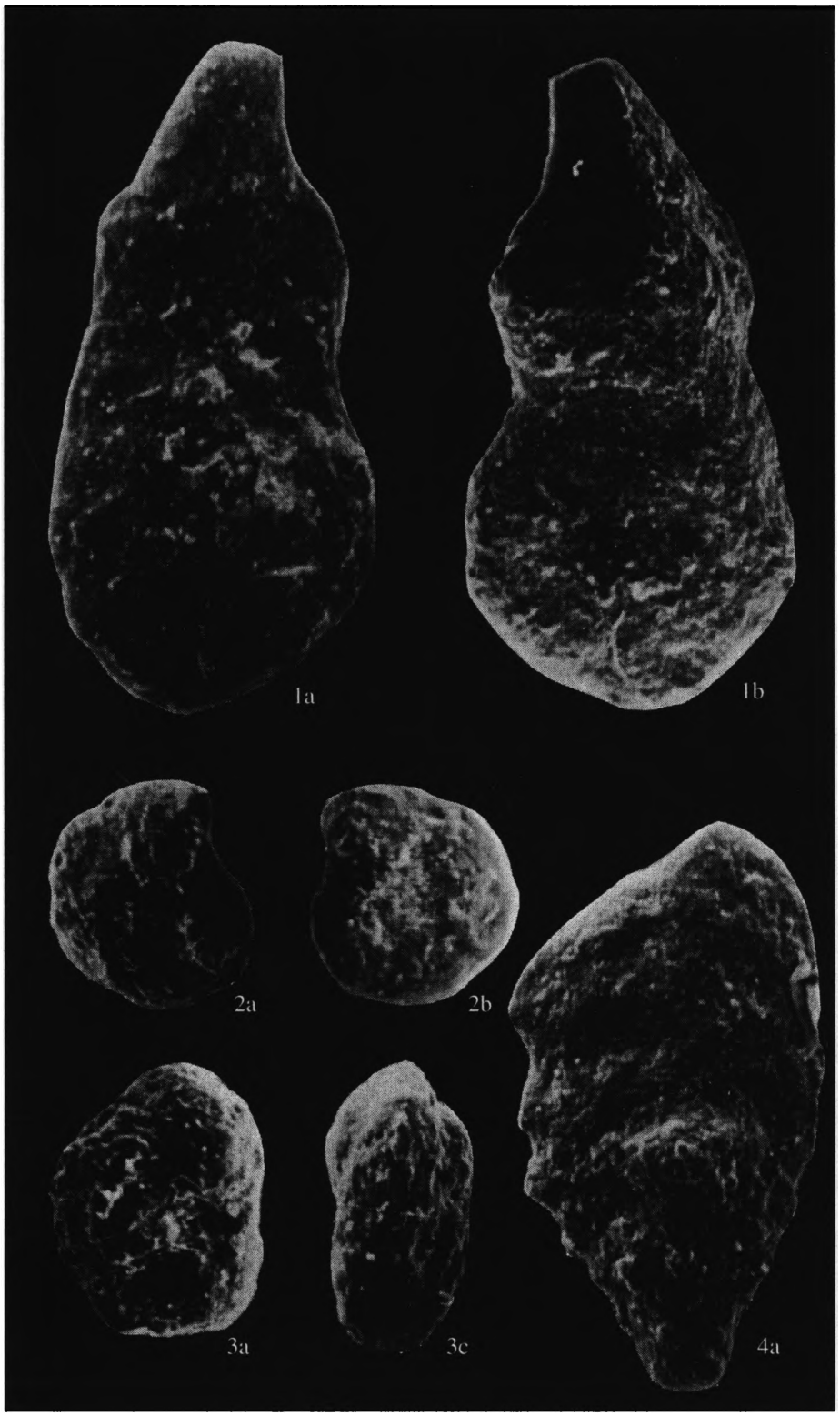

Plate II

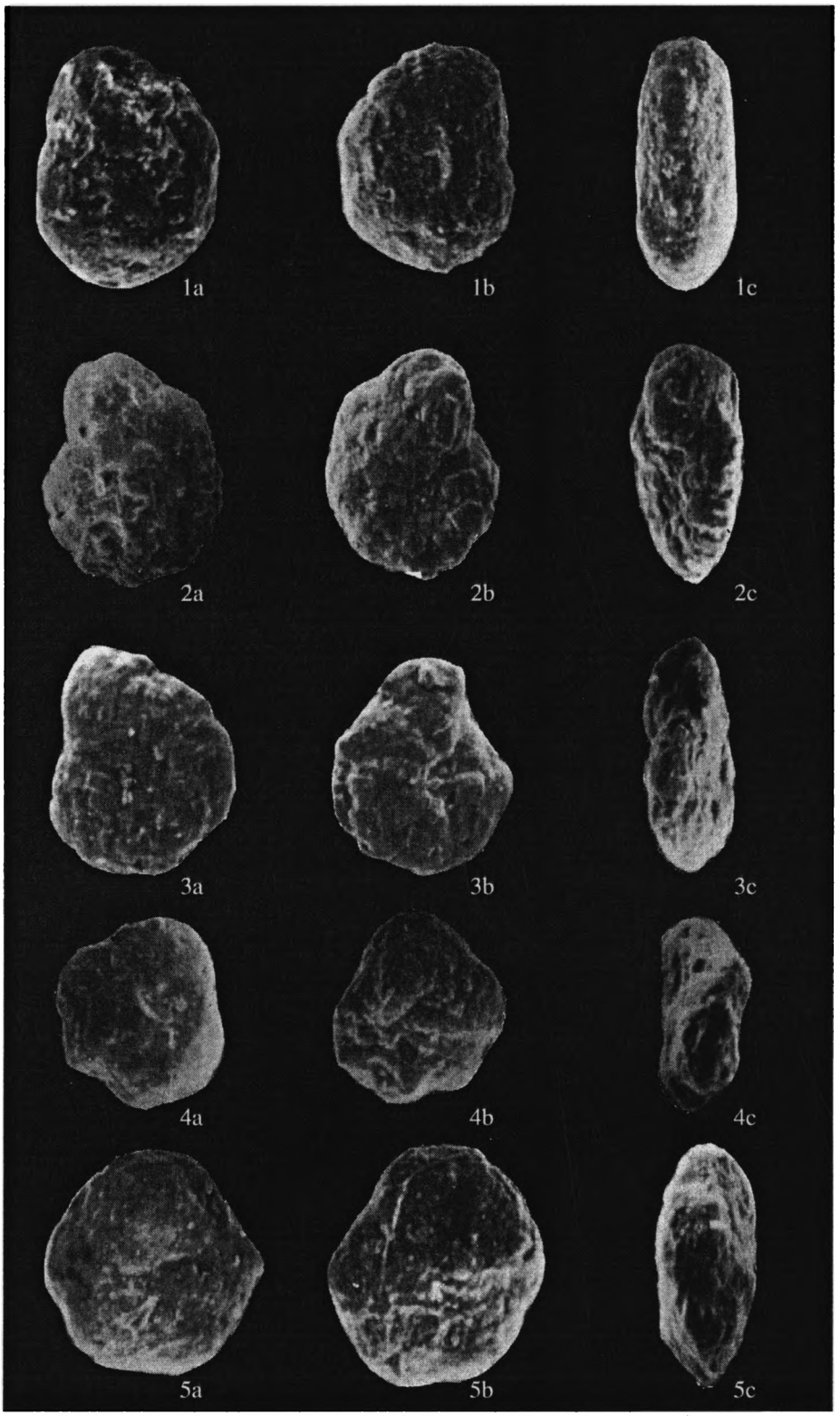

Plate II. Foraminifers from the lower Eocene deposits of West Siberia: (a) dorsal view; (b) ventral view; (c) aperture view; magnification ×100 for all foraminifers.

(1) *Labrospira granulosa* (Lipman), sample no. 1671 borehole (BH) 1, depth 460 m, Ypresian; (2) *Ammomarginulina spectata* Podobina, holotype no. 1543, BH 1, depth 450 m, Ypresian; (3) *Textularia carinatiformis* (Morosova) subsp. *sibirica* Podobina, paratype no. 1541, BH 1, depth 460 m, Ypresian; (4) *Anomalinoides ypresiensis* (ten Dam), sample no. 1530, BH 1, depth 450 m, Ypresian; (5) *Planorotalites planoconicus* (Subbotina), sample no. 1548, BH 1, depth 470 m, Ypresian.

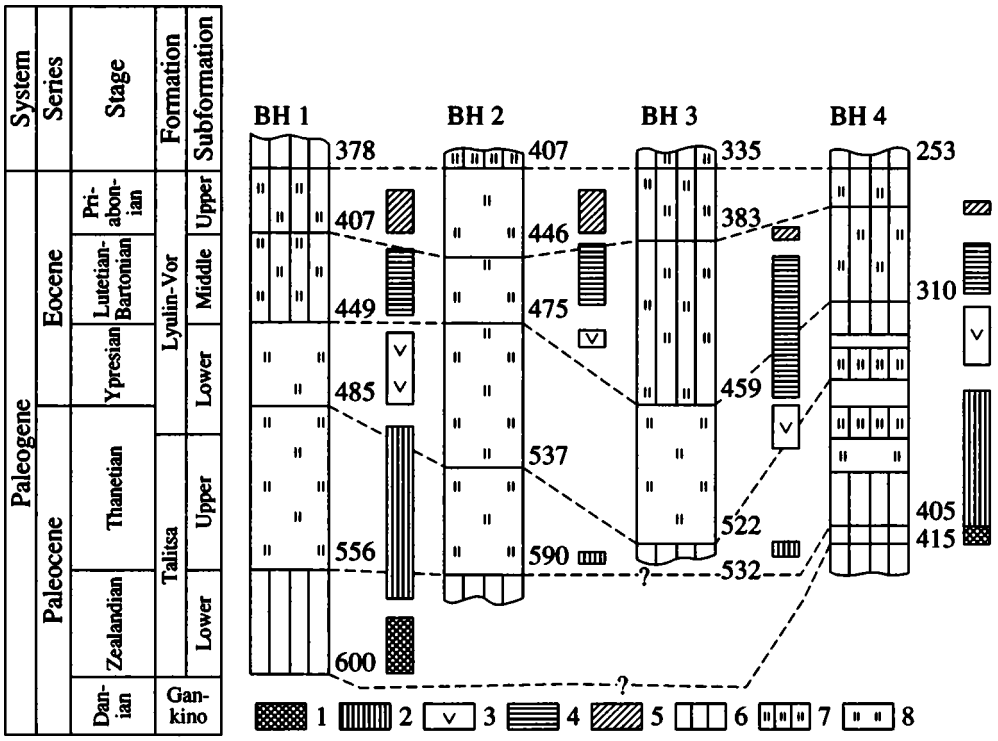

Fig. 2. Correlation of the Paleogene microfaunal zones in the Kaimysov arch sequences (boreholes 1, 2, 3, and 4 drilled in the basin of the north–south trending course of the Vasyugan River, Zapadnaya Group of the Tomsk Geological and Exploration Expedition): (1) *Ammoscalaria friabilis* Zone; (2) *Cibicidoides favorabilis* Beds (*Glomospira gordialiformis–Cyclammina coksuvorovae* Beds); (3) *Anomalinoides ypresiensis* Zone; (4) *Gaudryinopsis subbotinae* Zone; (5) *Labrospira honesta* Zone; (6) clay; (7) opoka-like clay; (8) opoka.

forms, the samples taken here at a depth of 470 and 450 m yielded calcareous benthonic and planktonic foraminifers.

The sample from the depth of 470 m contains diverse planorotaliids of the *Planorotalites pseudoscitulus* (Glaessner) and *P. planoconicus* (Subbotina) group typical of the upper half of the lower Eocene. Their presence allows one to assume that the proper layers correspond to the *Morozovella aragonensis* Zone.

Higher in the borehole section 1, the sample taken at a depth of 450 m marks a change in the species composition of foraminifers, and their assemblage comprises *Psammosphaera laevigata* White, *Reophax subfusiformis* Earland, *Labrospira granulosa* (Lipman), *Ammomarginulina spectata* Podobina, *Textularia carinatiformis* (Morosova) *sibirica* Podobina, *Anomalinoides nobilis* Brotzen, *Anomalinoides ypresiensis* (ten Dam), *Acarinina pentacamerata* (Subbotina), and others.

Representatives of the *Textularia* and *Acarinina* genera are as abundant in the sample as dozens of specimens. The nature and species composition of benthonic foraminifers suggest that the sample belongs to the uppermost lower Eocene. This is evident from the presence of the distinctive subspecies *Textularia carinatiformis* (Morosova) *sibirica* Podobina, whose more typical specimens are abundant at a higher level of the middle Lyulin-Vor Subformation of West Siberia. The *Anomalinoides ypresiensis* (ten Dam) species found at the depth of 470 and 450 m provides grounds to correlate, on the basis of benthonic calcareous foraminifers, the distinguished *Anomalinoides ypresiensis* Zone with the *Anomalinoides ypresiensis–Gaudryina hiltermanni* Zone of the lower Eocene deposits in the Netherlands (Doppert and Neele, 1983).

The foraminiferal assemblage with characteristic species *Anomalinoides ypresiensis* (ten Dam) and *Acarinina pentacamerata* (Subbotina) was also found in the borehole section 2 (at a depth of 479–474.9 m, Zapadnaya Group, the Kaimysov arch). They are similar to species from the borehole section 1, which are encountered at a depth of 450 m and indicate the early Eocene age of the deposits (the uppermost lower Eocene). The rocks of the studied interval also yielded abundant and diverse radiolarians of the early Eocene (Lipman, 1979).

In other areas of West Siberia (beyond the limits of the Kaimysov arch), the upper layers of the lower Lyulin-Vor Subformation assigned to the lower Eocene (gray opoka-like clays and opokas) usually contain singular poorly preserved planorotaliids, agglutinated foraminifers, and radiolarians. The *Ammosphaeroidina sphaerica* assemblage is locally encountered in the upper layers of the subformation. I found this assem-

blage previously in the borehole section 526 (the depth of 334 m) located near the village of Zav'yalovo in southern West Siberia. The assemblage was also encountered in the borehole 13n (the depth of 323–315 m; the Nyurolka River basin). The *Ammosphaeroidina* genus usually appears in the early Eocene. In the eastern area, gray opoka-like clays of this stratigraphic level bear rare and poorly preserved foraminifers of the *Psammosphaera, Reophax, Haplophramoides, Adercotryma,* and *Planorotalites* genera and also single radiolarians of the *Cenosphaera, Cenodiscus, Stylotrochus* and other genera along with their pyritized casts. The deposits are assigned arbitrarily (by the position in the section) to the lower Eocene and correspond to the *Anomalinoides ypresiensis* Zone.

Higher in the sequence, there is the *Gaudryinopsis subbotinae* Zone, which is described in my previous works (Podobina, 1975, 1984) and assigned to the middle Eocene (Lutetian–Bartonian) of western Siberia. The upper Eocene (Priabonian) is established in the range of the *Labrospira honesta* Zone (the upper Lyulin-Vor Subformation). The Priabonian age of this zone is confirmed by the planktonic *Globigerina officinalis* Subbotina species found along with agglutinated forms in the Chizhapka River basin (the borehole 5k, depth interval of 242–235 m). More detailed information on the above-mentioned zones and controversial topics related to this part of the sequence was presented in my previous publication (Podobina, 1990).

The aforementioned middle and upper Eocene zones have been recently revealed in boreholes 1, 2, 3, and 4 drilled within the central area of the Kaimysov arch (Fig. 2).

CONCLUSION

The performed analysis of foraminifers from western Siberia and their correlation with similar forms from adjacent provinces revealed the common species of benthonic calcareous and planktonic microfauna in the Paleocene and early Eocene paleocoenoses occurring in the studied region—Turgai and Western Europe, the type sections included. The correlation of West Siberian zones with coeval zones in the Turgai and European stratotypes of the stages, which is based on similarity of foraminiferal paleocoenoses, made it possible to establish zones of benthonic foraminifers in Western Siberia and to correlate them with planktonic foraminiferal zones of the oceanic scale.

The boundary between the upper Paleocene and lower Eocene, in other words, the boundary between the Thanetian and Ypresian, can be placed between the *Cibicidoides favorabilis* and *Anomalinoides ypresiensis* zones.

Reviewers V.A. Krasheninnikov and I.A. Basov

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Oligocene and Lower Miocene Diatom Zonation in the North Pacific

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Abstract—This article presents the Oligocene and lower Miocene diatom zonation for the North Pacific. The data obtained in Leg 145 of the Oceanic Drilling Program (ODP) allowed a subdivision of Oligocene and lower Miocene deposits of the subarctic Pacific into seven zones. In addition, the correlation between diatom assemblages from the Komandorskie and Japanese Islands and the high southern latitudes enabled the first recognition of a lower Oligocene zone in the North Pacific (the *Rhizosolenia oligocaenica* Zone with the lower and upper boundaries at the levels of 33.7 and 30.2 Ma, respectively). Diatom assemblages from more than 20 sections of middle–high latitudes of the Pacific region (Kamchatka, Sakhalin, Japan, and other areas) are correlated with the suggested zonation.

Key words: *Oligocene, lower Miocene, North Pacific, biostratigraphy, correlation, diatoms, zonation*

INTRODUCTION

The subdivision and correlation of Oligocene and lower Miocene deposits present one of the most complicated problems in the North Pacific region. At present, the refined diatom zonation has been elaborated and widely used for subdividing the Neogene deposits in the middle–high latitudes of the North Pacific, beginning from their middle Miocene beds (Akiba, 1986; Koizumi and Tanimura, 1985; Barron and Gladenkov, 1995). The zonation makes it possible to correlate the Upper Cenozoic marine deposits inside this vast territory. Evident progress has also been achieved in the Paleocene and Eocene zonations based on planktonic foraminifers and nannoplankton (Gladenkov *et al.*, 1988; Krasheninnikov *et al.*, 1988, 1989; Ben'yamovskii *et al.*, 1992; Okada and Kaiho, 1992; Volobueva *et al.*, 1994; Beaufort and Olafsson, 1995). In contrast, the Oligocene zones and their boundaries in the North Pacific are still poorly substantiated, because, on the one hand, data on relatively complete and sufficiently fossiliferous sequences have not been available up to present and, on the other hand, the carbonate microplankton remains of this age are virtually missing in these latitudes.

Therefore, the latest data on marine fossil diatoms, which are obtained by deep-sea drilling in the North Pacific and from sections of its western framing structures, are of great significance. These data and previously published materials served as a basis for the suggested stratigraphic subdivision and correlation of Oligocene and lower Miocene deposits in the North Pacific.

All radiological age estimates are presented in accordance with the geochronological scale of Cande and Kent (1992), where the latest radiometic and paleomagnetic information is taken into account. The dates for boundaries between series and subseries are also accepted according to this scale, particularly the dates of 33.7 Ma[1] for the Eocene–Oligocene boundary, 28.5 Ma for the lower–upper Oligocene boundary, 23.8 Ma[2] for the Oligocene–Miocene boundary, and 16.3 Ma for the lower–middle Miocene boundary.

MAIN RESULTS

Oligocene–Lower Miocene Zonation

Whereas the refined diatom zonations for the Oligocene and lower Miocene of the low latitudes of the North Pacific and the high latitudes of the Southern Ocean and North Atlantic are already elaborated and, to a certain extent, calibrated on the basis of magnetostratigraphy and carbonate plankton zonation (Barron, 1983, 1985; Fenner, 1984, 1985; Baldauf and Barron, 1991; Harwood and Maruyama, 1992; and others), the same for the middle and high latitudes of the North Pacific has just started.

The Oligocene and lower Miocene zonation in the North Pacific based on diatom assemblages from the Komandorskie and Japanese Islands and also on data

[1] This date coincides with that for the Eocene–Oligocene boundary of the Massignano section of Italy (Odin *et al.*, 1991) chosen as the boundary stratotype by the International Subcommission on Paleogene Stratigraphy (*The Eocene–Oligocene...*, 1988).

[2] This date coincides with this boundary age in the Lemme-Carrosio section of Italy, which was suggested as the Oligocene–Miocene boundary stratotype by the International Working Group on the Paleogene–Neogene Boundary (Steininger *et al.*, 1994).

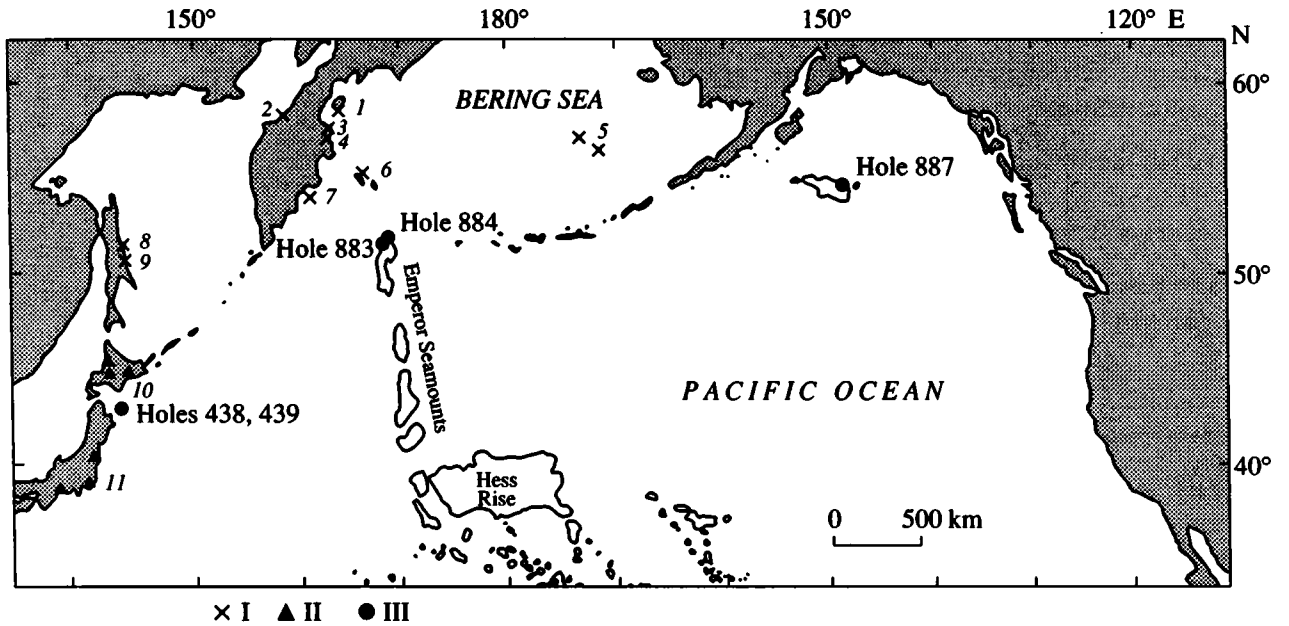


Fig. 1. Main localities of the Oligocene and early Miocene diatom assemblages in the North Pacific and its framing structures: (I) sections in the Komandorskie Islands, Kamchatka, Bering Sea, and Sakhalin: (*1*) Mys Tons, Il'khatunvayam, and Pestrotsvetnaya formations and *Laternula* Sandstone (Karaginskii Island); (*2*) Gakkh and Viventek formations (western Kamchatka, the Tochilo section); (*3*) Shagaevka Formation (Ozernoi Peninsula); (*4*) Ozernaya Formation (Cape Tupoi); (*5*) dredge samples from the Navarin basin of the Bering Sea; (*6*) Kamenka Formation (Bering Island); (*7*) dredge samples from the Kronotskii Gulf and Ol'ga Bay; (*8*) Mutnovka and Pilenga formations (Lun' depression); (*9*) Pilenga and Bora formations (Pogranichnyi depression); (II) sections of Japan (*10*, *11*); (III) ODP (883, 884, and 887) and DSDP holes (438 and 439)

obtained by Leg 145 of the Ocean Drilling Program (ODP) is discussed below.

(1) The Northwestern Pacific Ocean

Data obtained by ODP Leg 145 enabled Gladenkov and Barron (1995) to establish for the first time the upper Lower Oligocene to upper Lower Miocene diatom zonation in this region. Seven zones have been recognized according to successive first occurrences (FO) of zonal species in the virtually continuous diatom-bearing section of the ODP Hole 884B (Fig. 1) in the Detroit Seamount of the northwestern Pacific Ocean (51°27′026″ N, 168°20′228″ E; sea depth 3827 m; cores 63–71). Biostratigraphic levels of deposits younger than 18 Ma were calibrated on the basis of paleomagnetic data. In accordance with the scale of Cande and Kent (1992), the lower boundaries of zones (and, correspondingly, the first occurrences of zonal species) were dated as follows (from the base upward): the *Rocella vigilans* Zone as 30.2 Ma; the *Cavitatus rectus* Zone as 29.6 Ma; the *Rocella gelida* Zone as 28.2 Ma; the *Thalassiosira praefraga* Zone as 24.0 Ma; the *Thalassiosira fraga* Zone as 20.3 Ma; the *Cruciden-ticula sawamurae* Zone as 18.4 Ma; and the *Cruciden-ticula kanayae* Zone as 16.9 Ma (Fig. 2). The Oligocene–Miocene boundary (23.8 Ma) lies near the base of the *Thalassiosira praefraga* Zone ranging from 24.0 to 20.3 Ma. It should be noted that the successive lower Miocene diatom zones beginning from that of *Thalassiosira fraga* are also recorded in the ODP Hole 887, Gulf of Alaska (54°21′9″ N, 148°26′8″ W, sea depth 3645 m), and Hole 883, the Detroit Seamount (51°11′9″ N, 167°46′1″ E, sea depth 2396 m; Figs. 1, 2; Barron and Gladenkov, 1995).

Thus, the upper lower Oligocene and lower Miocene of the subarctic North Pacific are now subdivided into zones with age ranges from 0.6 to 4.2 Ma. However, the lower Oligocene zonation is not yet elaborated.

Problem of the lower Oligocene subdivision. Great progress has been recently achieved in the diatom-based calibration of the lower Oligocene, especially for the high latitudes of the Southern Ocean. As a result of deep-sea drilling in the Kerguelen Plateau and Prydz Bay of the East Antarctic coast (ODP Legs 119 and 120; Baldauf and Barron, 1991; Harwood and Maruyama, 1992) and in the Ross Sea shelf (MSSTS-1 and CIROS-1 projects; Harwood, 1986, 1989; Harwood *et al.*, 1989), the previous diatom zonations for the high southern latitudes have been significantly modified (Gombos and Ciesielski, 1983; Fenner, 1984, 1985) and zonal and infrazonal boundaries have been more precisely dated. The paleomagnetic calibration of biotic datum levels in these sections provides the basis to refine the radiological age values for the first (FO) and last occurrence (LO) diatom forms important in stratigraphy. As a result, the refined zonations, particu-

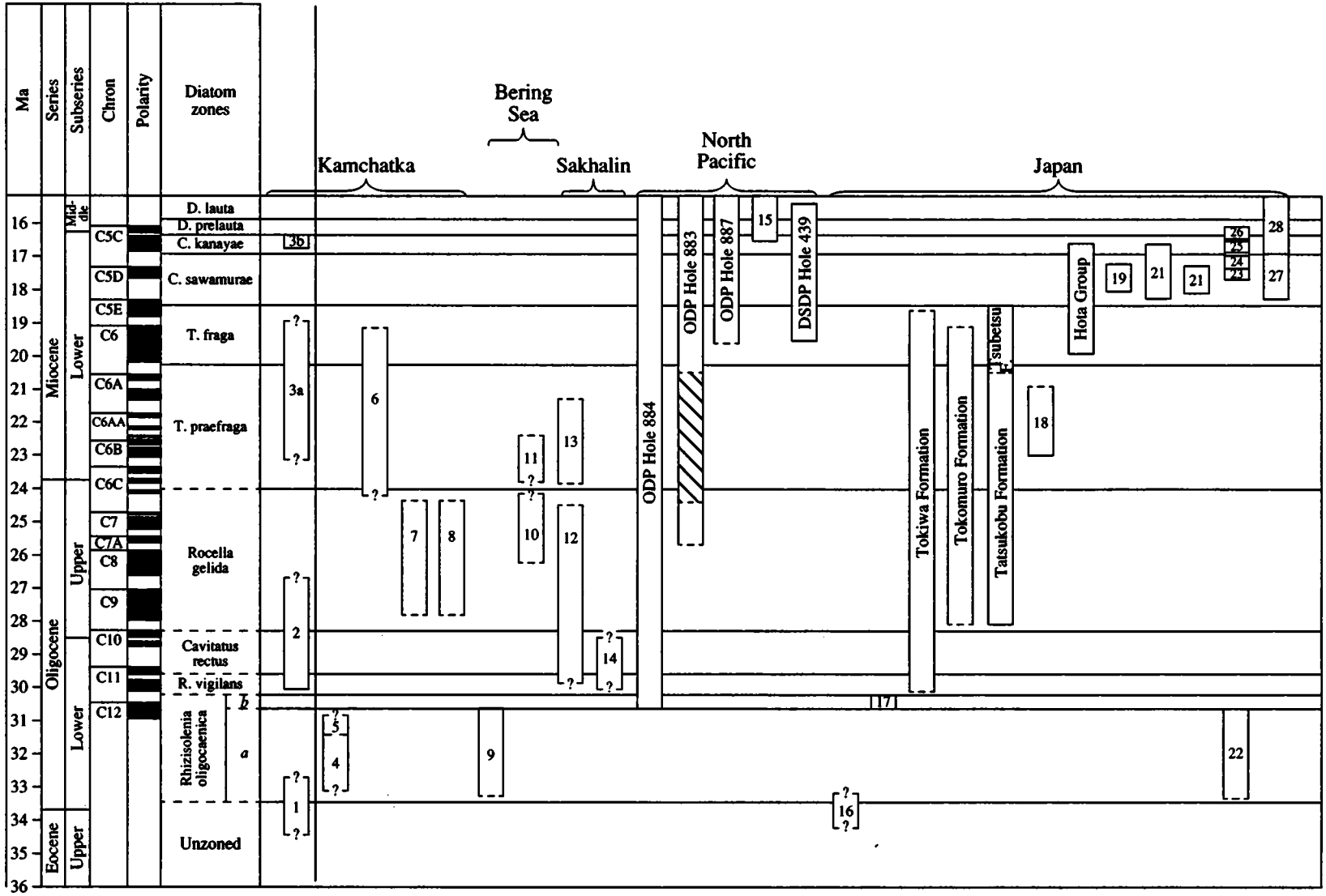


Fig. 2. Correlation of the North Pacific sections and the composite diatom zonation for the North Pacific (units from the *Rocella vigilans* to *Denticulopsis praelauta* zones are after Gladenkov and Barron, 1995; the *Rhizosolenia oligocaenica* Zone is proposed in this article): (*1*) Mys Tons Formation; (2) Il'khatunvayam Formation and *Laternula* Sandstone; (*3*) lower (a) and upper (b) parts of the Pestrotsvetnaya Formation; (4) Gakkh Formation; (5) Viventek Formation; (6) Ozernaya Formation; (7) Shagaevka Formation; (8) dredged deposits of Ol'ga Bay and Kronotskii Gulf; (9) Kamenka Formation; (10, 11) dredged deposits of the Navarin basin, Bering Sea; (12) Pilenga Formation of the Pogranichnyi depression; (13) the lower part of the Bora Formation; (14) Mutnovka and Pilenga formations of the Lun' depression; (15) DSDP Hole 438; (16) the upper part of the Poronai Formation; (17) Morawan Formation; (18) the upper part of the Sankebetsu Formation; (19) Yamanouchi Formation; (20) Morozaki Group; (21) Ikuchise Group; (22) Shirasaka Formation; (23) Kamenoo Formation; (24) Taira Formation; (25) Shirado Group; (26) Takaku Group; (27) Nabuto Formation; (28) Emi Formation.

larly for the lower Oligocene of the high latitudes of the Southern Ocean, were suggested (Baldauf and Barron, 1991; Harwood and Maruyama, 1992).

The Oligocene diatom zonation for the high latitudes of the southern hemisphere facilitates the subdivision of coeval deposits in the North Pacific, because many zonal species and other forms important for the lower Oligocene stratigraphy of the Antarctic region are also found, as shown below, in some sections of the western framing structures of the North Pacific. Some valuable forms common for these regions are known from the high latitudes of the Atlantic Ocean. However, direct correlation with the North Atlantic zonal scale is difficult, because this scale mainly reflects the regional evolution of flora (the Norwegian Sea; Schrader and Fenner, 1976; Fenner, 1985; *Istoriya...*, 1979) and its zones are partly outlined according to the scale for the low latitudes (the Labrador Sea; Baldauf and Monjanel, 1989). In addition, the Oligocene sequences of these regions lack adequate diatom characteristics and include many hiatuses. Moreover, the biotic datum levels are not calibrated here on the magnetostratigraphic basis.

The most interesting North Pacific section, characterized by well preserved and diverse Oligocene diatoms, is located in Bering Island (the Komandorskie Islands, Fig. 1). Diatom remains are found here in the upper Kamenka Formation (about 150 m thick) and composed of diatomites and tuffaceous diatomites. The diatom assemblages from this formation were previously considered as the early Oligocene–earliest late Oligocene in age (Gladenkov, 1990; Gladenkov, 1991), and these dates were coordinated with the silicoflagellate distribution (Gladenkov, 1992). However, the recently obtained data called in question the correctness of dating of the Komandorskie flora.

The latest investigations made it possible to suggest two versions of lower Oligocene diatom zonation for the high southern latitudes (Baldauf and Barron, 1991; Harwood and Maruyama, 1992). Both versions include the diatom *Rhizosolenia oligocaenica* Zone (Fig. 3). Harwood and Maruyama (1992) suggested its base to be defined by the FO *Rhizosolenia oligocaenica* Schrader (=*Rhizosolenia gravida* Gombos et Ciesielski) at the level of 33.3 Ma (C13N—hereinafter the radiological dates are followed by numbers in parentheses of magnetic polarity chrons and subchrons according to Berggren *et al.*, 1985) and its top by the FO *Cavitatus jouseanus* (Sheshukova-Poretskaya) Williams (=*Synedra jouseana* Sheshukova-Poretskaya) at 30.6 Ma (C12N). This zone consists of three subzones. Subzone "a" spans the interval from the base to the LO *Eurossia irregularis* (Greville) Sims (=*Triceratium polymorphus* Harwood et Maruyama =*Tricetatium macroporum* Hajos sensu Gombos and Cieiwlski, 1983; Gladenkov, 1990; Baldauf and Barron, 1991) at the level of 32.2 Ma (C12R). Subzone "b" ranges from the LO *Eurossia irregularis* to LO *Rhizosolenia oligocaenica* at the level of about 31.0 Ma (C12R-1), and Subzone "c" extends from the LO *Rhizosolenia oligocaenica* to the FO *Cavitatus jouseanus* (Fig. 3). The top of the overlying *Synedra jouseana* (*Cavitatus jouseanus*) Zone is defined by the FO *Rocella vigilans* Fenner corresponding to 30.2 Ma (C11R-2). The other version (Baldauf and Barron, 1991) shows the *Rhizosolenia oligocanica* Zone as the interval between the first appearance of the index-species at the level of 33.7 Ma (C13R-2) and the first appearance of *Rocella vigilans* (the index-species of the overlying zone) at the level of 30.9–30.7 Ma (C11-C12). The zone is subdivided by the LO *Rhizosolenia oligocaenica* at the level of 31.3 Ma (C12R) into two subzones ("a" and "b"). Thus, the principal discrepancy between these versions concerns the first appearance levels of two species: *Cavitatus jouseanus* and *Rocella vigilans*. The first version shows the FO *Cavitatus jouseanus* and *Rocella vigilans* at the levels of 30.6 Ma (C12N) and 30.2 Ma (C11R-2), respectively. In the sections of the Ross Sea glacial shelf, the FO *Cavitatus jouseanuc* is recorded at the level of 31.0 Ma (Harwood, 1986; Harwood *et al.*, 1989). The appearances of these two species at different levels resulted in establishing the *Synedra* (*Cavitatus*) *jouseana* Zone (Harwood and Maruyama, 1992; Harwood *et al.*, 1989). According to the second version (Fig. 3), the FO *Cavitatus jouseanus* and FO *Rocella vigilans* is within the intervals of 30.7–30.4 Ma (C12N) and 30.9–30.7 Ma (C11-C12), respectively (Baldauf and Barron, 1991). Because these intervals partly overlap each other, its authors recognized only the *Rocella vigilans* Zone above the *Rhizosolenia oligocanica* Zone and defined the boundary between them by the FO *Rocella vigilans* at the level of 30.9–30.7 Ma. Both species (*Cavitatus jouseanus* and *Rocella bivilans*) are absent from the assemblages of the Komandorskie Islands. However, the assemblages of the Kamenka Formation contains, along with abundant representatives of genera *Pyxidicula* (*Stephanopyxis*), *Hemiaulus*, and *Coscinodiscus*, the following forms: *Rouxia obesa* Schrader, *R. granda* Schrader, *Rhizosolenia oligocaenica* Schrader, *Rh. antarctica* Fenner, *Eurossia irregularis* (Greville) Sims, *Asteromphalus oligocenicus* Schrader et Fenner, *Asterolampra schmidtii* Hajos, *Navicula udintsevii* Schrader et Fenner, *Sceptroneis pesplanus* Fenner et Schrader, *Coscinodiscus hajosiae* Fenner, *Pyxilla reticulata* Grove et Sturt, *P. gracilis* Tempere et Forti, *Hemiaulus incisus* Hajos, *Triceratium unguiculatum* Greville, and *Pseudotriceratium radiosoreticulatum* Grunow (Plates I, II). All these forms are of limited stratigraphic range and typical of the lower Oligocene *Rhizosolenia oligocaenica* Zone (present in both versions) of the high latitudes of the Southern Ocean. The first and last occurrence levels of nearly all listed species are confined to this zone in the Antarctic regions. Therefore, supposing the almost synchronous formation of the high latitude floras in the Southern and North Pacific oceans, we can assign the Kamenka Formation assemblages to the *Rhizosolenia oligocaenica* Zone,

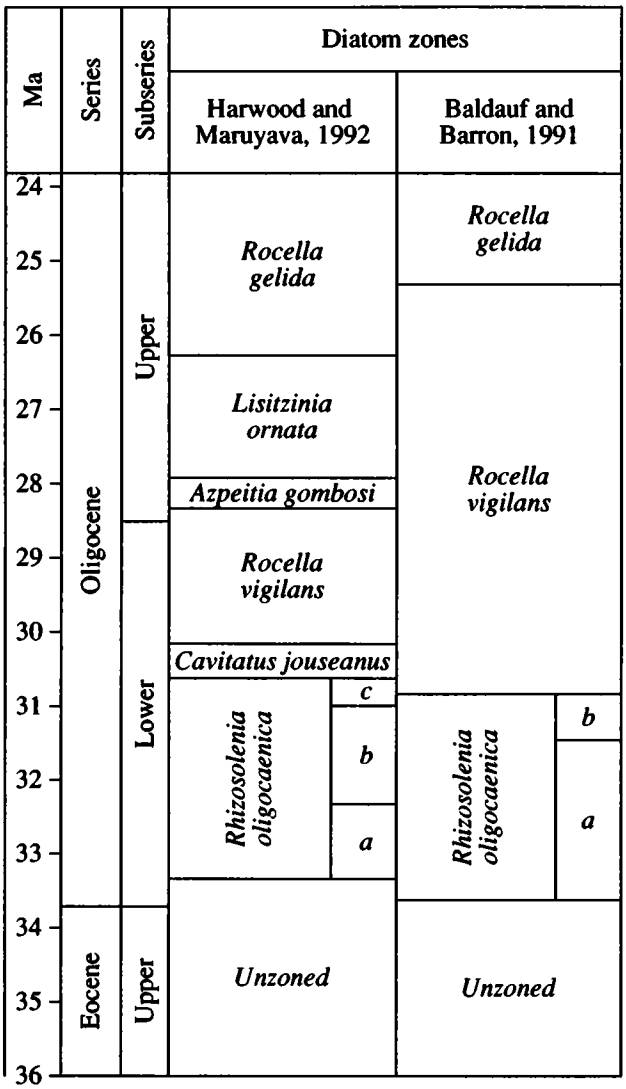


Fig. 3. The Oligocene diatom zonation for the East Antarctic coastal area (the Kerguelen Plateau and Prydz Bay). Radiological dates are from the scale of Cande and Kent (1992). *a, b, c* indicate subzones.

considering it either in the full scope, as interpreted by Harwood and Maruyama (1992), or without the uppermost horizons, according to Baldauf and Barron (1991). The enclosing deposits are dated back to the lower Oligocene range of 33.7–31.0 Ma. The subzonal specification is impossible because of some difference in radiological dates for the LO *Eurossia irregularis* (32.3 and 33.4–33.3 Ma) and LO *Rhizosolenia oligocanica* (31.0–30.9 and 31.4–31.3 Ma), although they were recovered from closely located holes of ODP Legs 119 and 120. The deep-sea drilling in the Labrador Sea (ODP Hole 647A and DSDP Hole 112) revealed that the first and last occurrence levels of some stratigraphically important species slightly differ from those in the high southern latitudes (Baldauf and Monjanel, 1989). For example, the synchronous LO *Rhizosolenia oligocaenica* and LO *Eurossia irregularis* (=*Ttinacria* sp. 1 sensu Baldauf and Monjanel, 1989, p. 344, Pl. 1, Fig. 10) are recorded in the north at the earlier level of ca. 30 Ma. The FO *Cavitatus jouseanus* coincident with the LO *Asterolampra schmidtii* in the Labrador Sea is at a lower level of 32.0 Ma. In the Southern Ocean, the LO *Asterolampra schmidtii* is dated as 30.6 Ma old (Baldauf and Barron, 1991). The FO *Rhizosolenia oligocaenica, Asterolampra schmidtii*, and *Asteromphalus oligocenicus* are almost synchronous in the North Atlantic and at the high southern latitudes (ca. 33.3–33.6 Ma). The age of the FO *Asterolampra schmidtii* is determined as 33.3–33.4 Ma in the Southern Ocean (Baldauf and Barron, 1991). The FO *Rouxia obesa* is recorded in the lower part of the *Rhizosolenia oligocaenica* Zone (C13N) in the southern hemisphere (Baldauf and Barron, 1991) and approximately at the same level (ca. 33.7 Ma) in the northern hemisphere (Baldfauf and Monjanel, 1989).

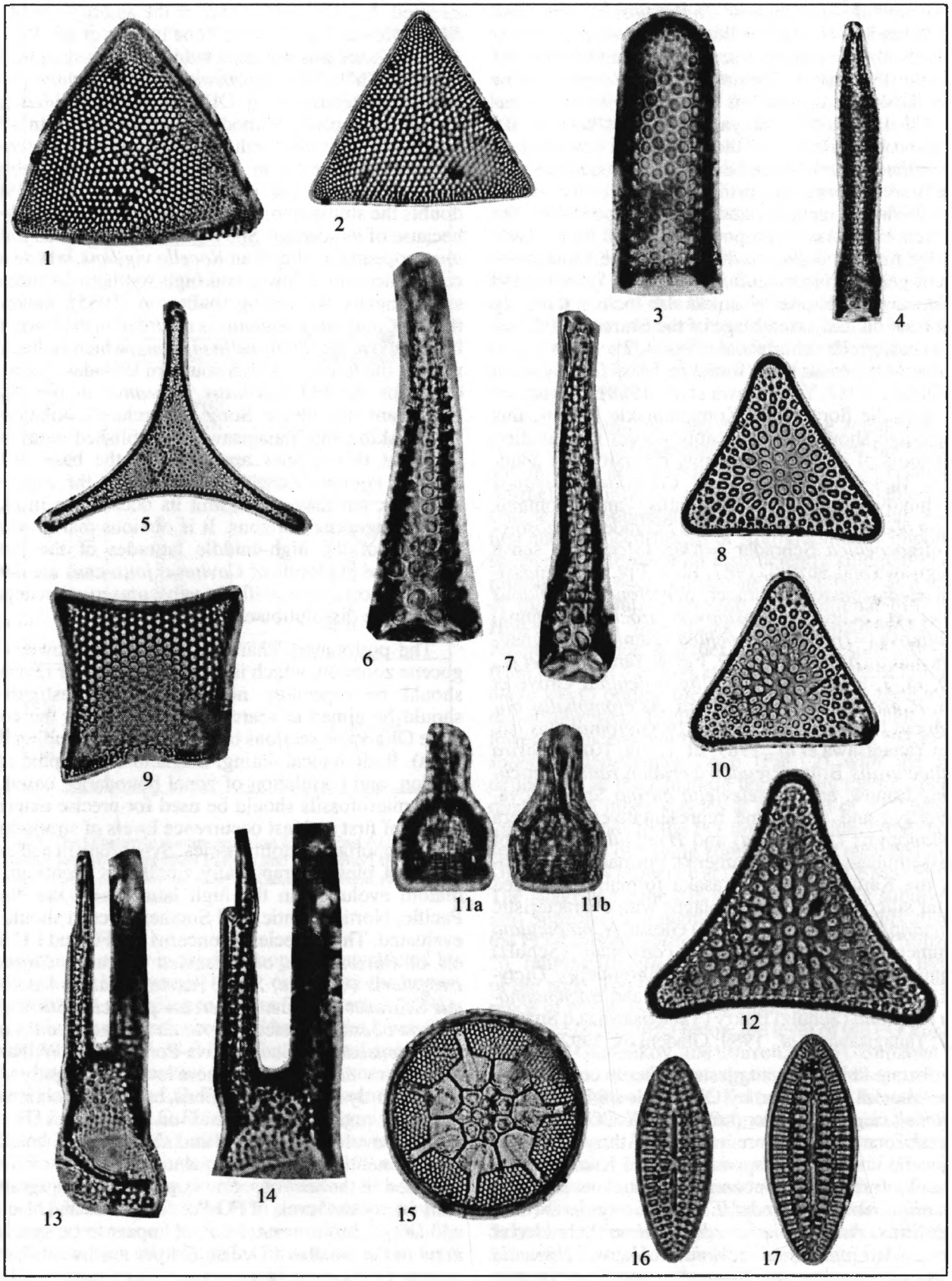
The diatom-based lower Oligocene zonation in the North Pacific is inadequately substantiated, because the datum levels are not reliably calibrated here relative to magnetostratigraphy, and there are no available lower Oligocene continuous sections to observe successive first and last occurrences of stratigraphically important species. The forms with insufficiently silicified valves missing from many Oligocene assemblages as a result of their dissolution presents an additional difficulty. However, data on sections of the Komandorskie and Japanese islands together with the results obtained from the ODP Site 884 and the high southern latitudes allow me to outline the Oligocene diatom zonation in the northwestern Pacific region. The lower half of the lower Oligocene can be recognized here as the *Rhizosolenia oligocaenica* Zone (Fig. 2) with the base defined, as in the Antarctic regions, by the first occurrence of the index-species at the level of 33.7 Ma. Its top corresponds to the FO *Rocella vigilans* (the index-species of the overlying zone), which is recorded in the ODP Site 884 sequence at the level of 30.2 Ma (Fig. 2). The suggested zone can be conventionally subdivided into Subzone "a" ranging from the base of this zone to the FO *Cavitatus jouseanus* (ca. 31.0 Ma) and Subzone "b" ranging from the FO *Cavitatus jouseanus* to the top of the zone. The last subdivision is based on the occurrence of rare *Cavitatus jouseanus* in the ODP Site 884 sequence (Core 74X) below the level of the FO *Rocella vigilans* (Gladenkov and Barron, 1995). Subzone "a" bears, first of all, the diatom assemblage of the Kamenka Formation (the Komandorskie Islands) consisiting of *Rouxia obesa* Schrader, *R. granda* Schrader, *Rouxia* spp., *Rhizosolenia oligocaenica* Schrader, *Rh. antarctica* Fenner, *Eurossia irregularis* (Greville) Sims, *Asteromphalus oligocenicus* Schrader et Fenner, *Asterolampra schmidtii* Hajos, *Navicula udintsevii* Schrader et Fenner, *Asteptroneis persplanus* Fenner et Schrader, *Coscinodiscus hajosiae* Fenner, *Pyxilla reticulata* Grove et Sturt, *P. gracilis* Tempere et Forti, *Hemiaulus incisus* Hajos, *H. rectus* var. *twista* Fenner,

Triceratium unguiculatum Greville, *Pseudotriceratium radiosoreticulatum* Grunow, *Thalassiosira mediaconvexa* Schrader, *Th. bukryi* Barron, *Melosira architecturalis* Brun, *Costopyxis trochlea* (Hanna) Strelnikova (=*Trochosira trochlea* Hanna), and *Kisseleviella carina* Sheshukova-Poretskaya, but lacking *Cavitatus jouseanus* (Sheshukova-Poretskaya) Williams (Plates I, II). The good preservation of the early Oligocene diatom flora in the Komandorskie Islands should be mentioned once more. It suggests an inconsiderable dissolution effect during diagenesis, and this is supported by the occurrence of forms with poorly silicified thin valves, e.g., by *Kannoa japonica* Komura, some representatives of genera *Rhizosolenia, Hemiaulus, Synedra,* and *Chaetoceros*. Subzone "a" could also include the early Oligocene diatom assemblage of the Shirasaka Formation in eastern Honshu Island (Figs. 4, 2), which is one of the oldest assemblages found in Japan (Yanagisawa and Suzuki, 1987; Yanagisawa *et al.,* 1989). In comparison with the flora of the Komandorskie Islands, this assemblage shows the noticeably poorer preservation of diatoms of a lower taxonomic diversity and abundance. However, it also lacks *Cavitatus jouseanus* (Sheshukova-Poretskaya) Williams and contains *Rouxia obesa* Schrader, *R. granda* Schrader, *Rhizosolenia oligocaenica* Schrader (=*Rhizosolenia* sp. sensu Yanagisawa and Suzuki, 1987, Pl. 2, Fig. 6), *Thalassiosira mediaconvexa* Schrader, *Sceptroneis pesplanus* Fenner et Schrader, *Costopyxis trochlea* (Hanna) Strelnikova (=*Trochosira trochlea* Hanna sensu Yanagisawa and Suzuki, 1987, Pl. 2, Fig. 4; Yanagisawa *et al.,* 1989, Pl. 7, Figs. 28–31), *Pyxilla reticulata* Grove et Sturt, *P. gracilis* Tempere et Forti, *Asteromphalus oligocenicus* Schrader et Fenner (=*Asteromphalus* sp. sensu Yanagisawa *et al.,* 1989, Pl. 7, Fig. 10), *Melosira architecturalis* Brun, *Pseudotriceratum radiosoreticulatum* Grunow, and *Kisseleviella carina* Sheshukova-Poretskaya and also some representatives of genera *Stephanopyxis* (*Pyxidicula*) and *Hemiaulus* typical of the assemblage from the Kamenka Formation. In addition, the Kamenka and Shirasaka formations yielded similar silicoflagellate assemblages with characteristic *Naviculopsis trispinosa* (Schulz) Glezer, *N. biapiculata* (Lemmermann) Frenguelli, *N. constricta* (Schulz) Frenguelli, *Corvisema triacantha* (Ehrenberg), *Dictyocha deflandrei* Frenguelli ex Glezer, and *Bachmannocena apiculata* (Schulz) Bukry (Yanagisawa and Suzuki, 1987; Yanagisawa *et al.,* 1989; Gladenkov, 1992).

Subzone "b" of the suggested scheme corresponds to the interval penetrated by ODP Hole 884 below the FO *Rocella vigilans* Fenner (samples 74X-CC–71X-CC). This subzone is free of forms typical of the older lower Oligocene interval corresponding to the Kamenka and Shirasaka formations (Subzone "a"), such as *Rhizosolenia oligocaenica* Schrader, *Eurossia irregularis* (Greville) Sims, *Asteromphalus oligocenicus* Schrader et Fenner, *Asteroplampra schmidtii* Hajos, *Navicula udintsevii* Schrader, and *Costopyxis trochlea* (Hanna) Strelnikova, but it includes *Cavitatus jouseanus* (Sheshukova-Poretskaya) Williams. However, it should be stressed that the replacement of the assemblage of the *Rhizosolenia oligocaenica* Zone by that of the *Rocella vigilans* Zone was not detected in a single section, and Subzone "b" of the *Rhizosolenia oligocaenica* Zone, which was recovered by ODP Hole 884, yielded only rare diatom valves of moderate preservation. In addition, the stratigraphic value of FO *Cavitatus jouseanus* and its synchronism in different regions are ambiguously estimated. For instance, Fenner (1984, 1985) doubts the stratigraphic significance of this datum level because of its scarcity. She argues that *Cavitatus jouseanus* appeared earlier than *Rocella vigilans*, whose first occurrence in the lower and high southern latitudes is simultaneous. According to Barron (1985), however, the FO *Cavitatus jouseanus* is recorded in the lower latitudes above the FO *Rocella vigilans*, which is diachronous in the lower and high southern latitudes. Accurate dates for the FO *Cavitatus jouseanus* in the North Pacific are unavailable. Some researchers (Akiba *et al.,* 1993; Akiba and Yanagisawa, unpublished data) suggest that this species appears near the base of the *Rocella vigilans* Zone (particularly, in the Japanese sections), but they also admit its occurrence in older lower Oligocene horizons. It is obvious that, in many sections of the high–middle latitudes of the North Pacific, the FO levels of *Cavitatus jouseanus* are determined in accordance with the valve preservation dependent on the dissolution effect.

The preliminary character of the early lower Oligocene zonation, which is a subject of further revision, should be especially noted. Further investigations should be aimed at searching and studying the complete Oligocene sections bearing fossils (including diatoms). Radiological dating, magnetostratigraphic calibration, and correlation of zonal boundaries based on other microfossils should be used for precise determination of first and last occurrence levels of stratigraphically important diatom species. Synchronism and similarity of biostratigraphically significant events in the diatom evolution in the high latitudes of the North Pacific, North Atlantic, and Southern Ocean should be evaluated. This especially concerns the FO and LO levels of *Rhizosolenia oligocaenica* Schrader, *Eurossia irregularis* (Greville) Sims, *Asteromphalus oligocenicus* Schrader et Fenner, *Asterolampra schmidtii* Hajos, *Rouxia obesa* Schrader, *R. granda* Schrader, and *Cavitatus jouseanus* (Sheshukova-Poretskaya) Williams. Recent practice shows that these levels are slightly variable not only in different regions, but also in sections of a single region and depend on differences in the regional evolution of floras and also on sedimentation environments and subsequent diagenetic effects recorded in the sections. For example, the stratigraphically important levels of FO *Rocella gelida* and also FO and LO *Lisitzinia ornata* do not appear to be synchronous in the Southern Ocean and the northwest Pacific region. In these regions, they are dated respectively as 26.3, 27.9, and 24.3 Ma (Harwood and Maruyama,

Plate I



1992) and as 28.2, after 28.2, and ca. 20.3 Ma (Gladenkov and Barron, 1995). In addition, relations between the extinction time of many taxa and their LO levels in the sections are difficult to estimate due to the specific evolutionary features of different floras and the redeposition effect. For example, the LO *Pyxilla reticulata* is recorded at the level of 30.3 Ma (C11R-2) near the Antarctic coast (Harwood and Maruyama, 1992), but its distribution extends up to the lower Miocene in other sections of the high latitudes of the Southern Ocean and the northwestern Pacific (Fenner, 1985; Gladenkov and Barron, 1995). The FO levels seem to be more reliable, and some researchers prefer them to be used. Thus, every dated level should be used with some caution and controlled by other available information.

(2) The Northwestern Framing Structures of the Pacific Ocean

The suggested zonation facilitates correlation of the Oligocene and lower Miocene deposits in the entire North Pacific and its framing structures. However, the correlation of the majority of sections with the zonal scale is hampered by their fragmental paleontological characteristics and incomplete succession of stratigraphically important species. In addition, some sections even bearing the coeval Tertiary diatom floras show the varying "aspect" (taxonomic composition) of the latter dependent on environments of their formation (open ocean, neritic, and nearshore zones; different temperature and salinity regimes). Therefore, some diatom assemblages lack zonal species and other forms of stratigraphic value. Moreover, the lower Oligocene diatom assemblages are found, as a rule, in rare samples, and ages of the oldest Oligocene assemblages are inferred from their comparison with coeval assemblages from other regions (the Norwegian Sea, South Pacific, and others). Despite this, the correlation of Oligocene and lower Miocene deposits from Kamchatka and the Bering Sea in the north to Japan in the south becomes credible. The diatom assemblages from different areas of the North Pacific are comparatively analyzed below. Their localities are shown in Fig. 1 for the North Pacific and in Fig. 4 for Japan. Figure 2 presents correlation of these assemblages with the North Pacific diatom zonation.

Kamchatka. Diatom assemblages of the Oligocene age were identified in the 1980s in Kamchatka, mostly in its western and northeastern parts. However, the floral remains are detected here in rare samples, have a poor to moderate preservation state, and include few forms of stratigraphical value. Therefore, the inferred boundaries of biostratigraphic subdivisions are rather conventional. The Paleogene diatom assemblages are most representative in the section of Karaginskii Island, northeastern Kamchatka (Oreshkina, 1982; Gladenkov, 1991; Volobueva *et al.*, 1992). The oldest assemblage of this locality is confined to the upper part of the Mys Tons Formation. The species composition of the assemblage, including characteristic *Pyxilla reticulata* Grove et Sturt, suggests its latest Eocene–earliest Oligocene age. Foraminifers from the same part of the section indicate the late Eocene age (Krasheninnikov *et al.*, 1988). The second assemblage, which was found in the Il'khatunvayam Formation and overlying *Laternula* Sandstone, contains diverse representatives of genus *Pyxidicula* (*Stephanopyxis*), as well as single *Rocella vigilans* Fenner and *Paralia fausta* Schmidt. This impoverished assemblage can be assigned to the interval of the upper lower Oligocene–upper Oligocene (the *Rocella vigilans–Rocella gelida* Zone). The Oligocene age of the enclosing deposits is supported by foraminiferal records (Krasheninnikov *et al.*, 1988).

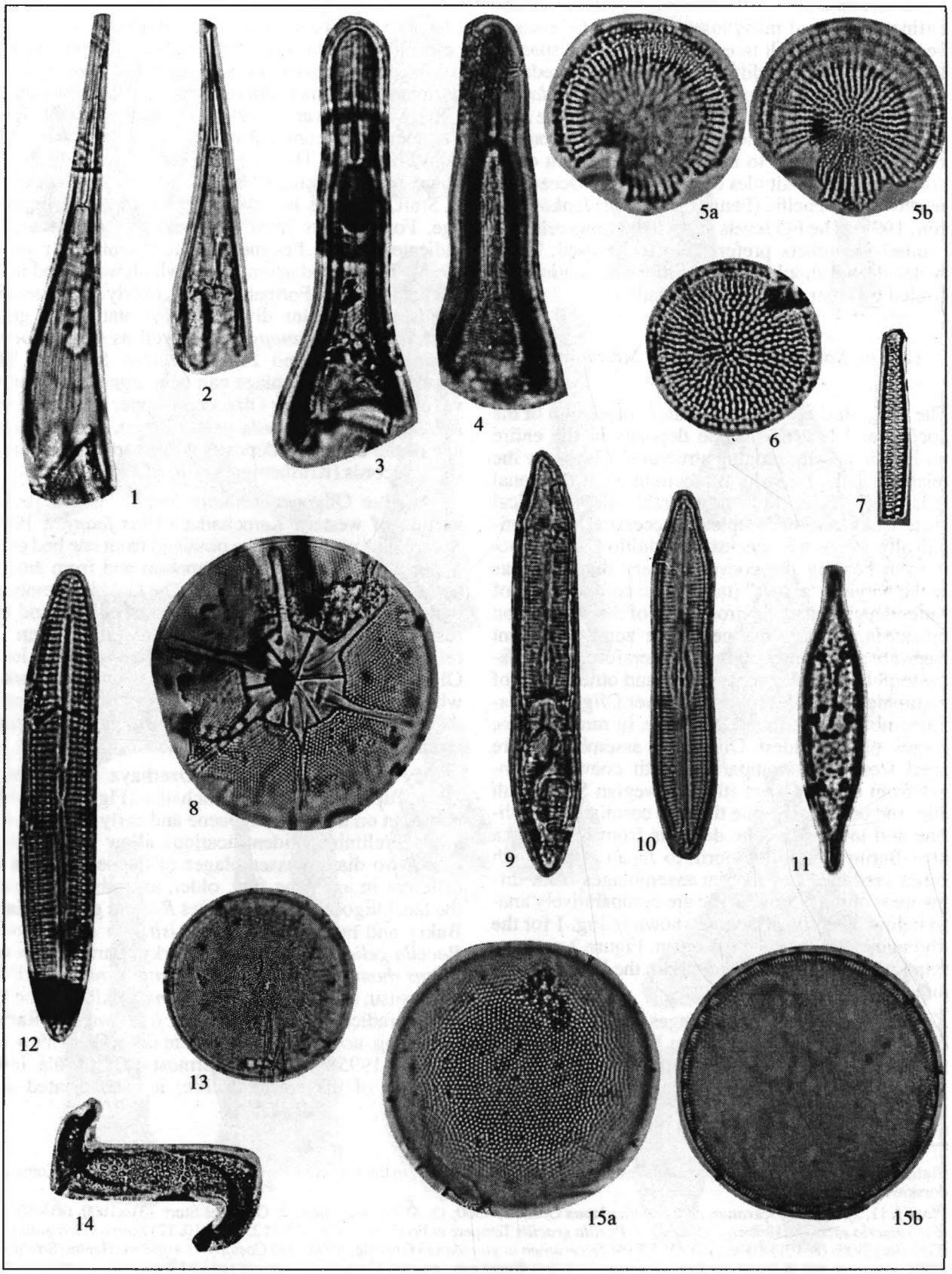
Another Oligocene diatom locality is the Tochilo section of western Kamchatka (*Atlas fauny...*, 1984). Scarce diatom remains are obtained from one bed of the upper part of the Gakkh Formation and from another bed of the Viventek Formation. The Gakkh assemblage including *Pyxilla* aff. *reticulata* Grove et Sturt and representatives of genus *Pseudotriceratium* can be referred to the Oligocene, most probably, to the lower Oligocene. The assemblage of the Viventek Formation, which is slightly different in composition, can also be dated as Oligocene in age. Both assemblages lack zonal species.

My investigation of the Ozernaya Formation in Cape Tupoi of eastern Kamchatka (Fig. 1) provided new data on the late Oligocene and early Miocene diatoms. Preliminary identifications allow me to distinguish two diatom assemblages of the section that are different in age. The first, older, assemblage includes the late Oligocene zonal species *Rocella gelida* (Mann) Bukry and rare specimens of *Lisitzinia ornata* Jouse, *Rocella gelida* var. *schraderi* (Burky) Barron, *Coscinodiscus rhombicus* Castracane, *Cavitatus rectus* Akiba et Hiramatsu, and *Kisseleviella ezoensis* Akiba. Some levels sporadically yielded *Actinocyclus ingens* Rattray appearing, according to the recent data (Gladenkov and Barron, 1995), in the lowermost part of the lower Miocene of the North Pacific; it is associated with

Plates I and II. Characteristic representatives of the diatom assemblage from the Kamenka Formation of Bering Island, the Komandorskie Islands

Plate I. (1, 2) *Pseudotriceratum radiosoreticulatum* Grunow, ×500; (3, 4) *Pyxilla reticulata* Grove et Sturt: (3) ×1000, (4) ×500; (5) *Trinacria excavata* Heiberg, ×500; (6, 7) *Pyxilla gracilis* Tempere et Forti, (6) ×1000, (7) ×1250; (8, 10, 12) *Eurossia irregularis* (Greville) Sims, (8, 10) ×1000, (12) ×1250; (9) *Triceratium unguiculatum* Greville, ×500; (11) *Costopyxis irochlea* (Hanna) Strelnikova, ×1250, a and b under different focuses; (13, 14) *Hemiaulus incisus* Hajos, (13) ×1000, (14) ×1250; (15) *Asterolampra schmidtii* Hajos, ×500; (16, 17) *Rouxia obesa* Schrader, ×1250.

Plate II



Rocella gelida (Mann) Burky at one level. This assemblage can be conventionally correlated with assemblages typical of the upper part of the *Rocella gelida* Zone and overlying the *Thalassiosira praefraga* Zone. The enclosing deposits can be assigned to the interval of the upper lower Oligocene to the lowermost Miocene. The second assemblage contains rare diatom forms, such as *Thalassiosira fraga* Schrader, *Actinocyclus ingens* Rattray, *Raphidodiscus marylandicus* Christian, *Rhizosolenia hotaense* Akiba, *Rh. norwegica* Schrader, and *Kisseleviella ezoensis* Akiba in association with silicoflagellate species *Naviculopsis lata* var. *obliqua* Bukry, but it lacks representatives of genus *Rocella*. It is correlative with the assemblage of the lower Miocene *Thalassiosira fraga* Zone of the North Pacific (Fig. 2). The enclosing deposits thus correspond to the first half of the lower Miocene. At present, the Ozernaya Formation is the first example of a microfossiliferous sequence in Kamchatka ranging from the upper part of the upper Oligocene to the lower part of the lower Miocene. The diatom assemblages from these deposits have been correlated to the North Pacific zonation based on data of ODP Leg 145. The significance of this correlation for the Cenozoic stratigraphy in Kamchatka consists in the possibility to correlate other assemblages, especially benthic fossils, with the diatom-based stratigraphic subdivisions.

The published data on diatoms from Ozernoi Peninsula of eastern Kamchatka (Litvinov *et al.*, 1990) also allows correlation of the distinguished assemblages with the North Pacific zonal assemblages. The diatom assemblage with *Rocella vigilans* and *Lisitzonia ornata* Jouse from the Shagaevka Formation can be correlated with that of the upper Oligocene *Rocella gelida* Zone. The same level is recorded in the bottom deposits at the southeastern Kamchatka coast. The dredge samples from submarine canyons of Ol'ga Bay and Kronotskii Gulf yielded the diatom assemblage of *Rocella vigilans* Fenner, *Lisitzinia ornata* Jouse, *L. ornata* f. *pentagona* Harwood, and *Sceptroneis tenue* Schrader et Fenner (Glezer *et al.*, 1986; Pushkar', 1987). This assemblage is also correlative to that of the abovementioned upper Oligocene zone of the North Pacific.

The late early Miocene diatom assemblage correlative with that of the *Crucidenticula kannayae* Zone was found only in the upper part of the Pestrotsvetnaya Formation of Karaginskii Island, northeastern Kamchatka (Gladenkov *et al.*, 1992). Below this level, the formation contains an older assemblage with *Actinocyclus ingens* Rattray and *Kisseleviella ezoensis* Akiba (Volobueva *et al.*, 1992), which can be correlated with the assemblages of the lower Miocene *Thalassiosira praefraga* and *Thalassiosira fraga* zones.

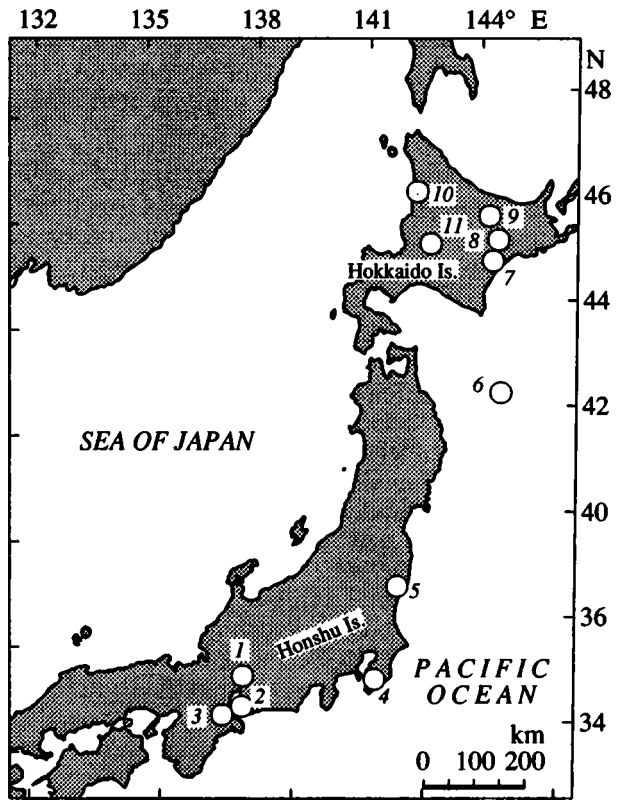


Fig. 4. Localities of the Oligocene and early Miocene diatom assemblages in Japan (according to Akiba, 1980, 1986; Akiba *et al.*, 1982, 1993; Akiba and Yanagisawa, unpublished data and personal communication): (*1*) Yamanouchi Formation; (*2*) Morozaki Group; (*3*) Ikuchise Group; (*4*) Hota Group, Nabuto, and Emi formations; (*5*) Kamenoo and Taira formations, Shirado and Takaku groups; (*6*) DSDP Holes 438 and 439; (*7*) Tokomuro Formation; (*8*) Morawan Formation; (*9*) Tatsukobu and Teubetsu formations; (*10*) the upper part of the Sankebetsu Formation; (*11*) the upper part of the Poronai Formation.

The Bering Sea. The Oligocene diatoms of the Bering Sea were recovered from samples dredged in the Navarin basin (Baldauf and Barron, 1987). Both of the distinguished diatom assemblages contain *Odontella sawamurae* Akiba (=*Odontella* sp. 1 sensu Baldauf and Barron, 1987, Pl. 3, Fig. 13, = *Odontella aurita* sensu Baldauf and Barron, 1987, Pl. 3, Figs. 7, 11) and *Kisseleviella ezoensis* Akiba (=*Cymatosira compacta* sensu Baldauf and Barron, 1987, Pl. 10, Fig. 3). A distinct feature of one of the assemblages is the presence

Plate II.

(1, 2) *Rhizosolenia oligocaenica* Schrader, ×1000; (3, 4) *Rhizosolenia antarctica* Fenner, ×1250; (5, 6) *Thalassiosira mediaconvexa* Schrader, ×1250, a and b under different focuses; (7) *Sceptroneis pesplanus* Fenner, ×500; (8) *Asteromphalus oligocenicus* Schrader et Fenner, ×1000; (9, 10) *Navicula udintsevii* Schrader, ×1250; (11) *Kisseleviella carina* Sheshukova-Poretskaya, ×1250; (12) *Rouxia granda* Schrader, ×1250; (13) *Coscinodiscus hajosiae* Fenner, ×500; (14) *Hemiaulus rectus* var. *twista* Fenner, ×500; (15) *Thalassiosira bukryi* Barron, ×500, a and b under different focuses.

of *Rocella gelida* (Mann) Bukry and *Thalassiosira mediaconvexa* Schrader. The assemblages can be conventionally correlated with flora of the upper Oligocene–lower Miocene *Rocella gelida* and *Thalassiosira praefraga* zones of the North Pacific.

Sakhalin. Early Cenozoic diatoms are known here from the Mutnovka, Bora, and Pilenga formations of the Pogranichnyi and Lun' depressions of eastern Sakhalin (Savitskii *et al.*, 1979; Boldyreva, 1991; Tusov *et al.*, 1991).

According to unpublished data of Boldyreva and the earlier paper of Savitskii *et al.* (1979), the upper part of the Pilenga Formation contains an assemblage of Oligocene age, which includes *Rocella vigilans* Fenner, *R. praenitida* (Fenner) Fenner, *Lisitzinia ornata* Jouse, *L. ornata* f. *pentagona* Harwood, and *Thalassiosira mediaconvexa* Schrader. The coexistence of these forms also characterizes the assemblage of the upper Oligocene *Rocella gelida* Zone, and both units seem to be correlative. The underlying deposits of the Pilenga Formation lacking zonal species most likely represent the undivided older Oligocene horizons. Some new data on diatoms from the lower part of the Bora Formation overlying the Pilenga Formation became available recently. These deposits yielded an assemblage consisting of *Odontella sawamurae* Akiba, *Cavitatus jouseanus* (Sheshukova-Poretskaya) Williams, *C. rectus* Akiba et Hiramatsu, *C. miocenicus* (Schrader) Akiba et Yanagisawa, *Ikebea tenuis* (Brun) Akiba, and *Kisseleviella carina* Sheshukova-Poretskaya associated with rare *Rhizosolenia miocenica* Schrader, *Actinocyclus ingens* Rattray, *Thalassiosira praefraga* Gladenkov et Barron, *Th. mediaconveca* Schrader, and *Lisitzinia ornata* Jouse. This assemblage can be correlated with that of the lower Miocene *Thalassiosira praefraga* Zone, and the enclosing deposits can be assigned to the lower half of the lower Miocene.

The diatom assemblage from the Mutnovka and Pilenga formations of the Lun' depression (Tuzov *et al.*, 1991; Boldyreva, 1991), as well as their forms from the lower part of the Pilenga Formation of the Pogranichnyi depression, have the neritic "aspect" and a poor to moderate preservation degree. Zonal species are missing among them, and their precise age is unknown. However, a certain similarity of this assemblage with the Japanese assemblages (the Morawan Formation) and, to some extent, with diatoms of the Komandorskie Islands indicates its Oligocene (most likely, the early Oligocene) age.

DSDP Holes 439 and 438. According to the International Deep-Sea Drilling Project (DSDP), these holes were drilled near the northeastern coast of Honshu Island, Japan. Before ODP Leg 145, they remained the only sites where the lower Miocene deposits bearing marine diatoms were recovered in middle–high latitudes of the North Pacific (Barron, 1980; Akiba *et al.*, 1982; Akiba, 1986; Yanagisawa, 1993). Cores 20-5 from Hole 439 (40°37′61″ N and 143°18′63″ E; sea depth 1656 m) yielded diatom assemblages corresponding to the interval of the *Thalassiosira fraga* (the upper part of the lower Miocene) and *Denticulopsis lauta* (the lower part of the middle Miocene) zones. Deposits corresponding to the *Crucidenticula sawamurae* Zone and a greater part of the *Crucidenticula kanayae* Zone are missing at the site due to the break in sedimentation (Akiba, 1986). The oldest diatom assemblage recovered from Holes 438A and 438B (40°37′79″ N and 14°14′15″ E; sea depth 1558 m) can be correlated with the assemblage of the *Crucidenticula kanayae* Zone of the upper part of the lower Miocene.

Japan. Some time ago, I collected samples from the upper part of the Poronai Formation, Hokkaido Island (Fig. 4), i.e., from the unit "i" according to Teshima (1962). Rare diatom valves of poor to moderate preservation were found for the first time in carbonate concretions of these horizons. In particular, rare *Pyxilla* spp.; *Hemiaulus* cf. *bipons* (Ehr.) Grun.; *H.* aff. *hostilis* Heiberg, Genus and sp. indet. Gombos, 1976; Genus and sp. indet. 1 Fenner, 1977 were preliminarily identified in this impoverished assemblage. It is difficult to date the flora precisely, but its composition suggests the late Eocene–late early Oligocene age. The sampled part of the Poronai Formation is assigned to the upper part of lower Eocene (Subzone CP 15b) by means of nannoplankton (Okada and Kaiho, 1992) and to the uppermost Eocene–lowermost Oligocene by planktonic foraminifers (Kaiho, 1983).

Some concretions from the Morawan Formation of eastern Hokkaido also yielded diatoms of the Oligocene age (Saito *et al.*, 1988). The assemblage comprises *Rouxia* spp., *Thalassiosira* cf. *bukryi* Barron, *Rhizosolenia antarctica* Fenner, *Pseudotriceratium radiosoreticulatum* Grunow, and various representatives of genus *Stephanopyxis* (*Pyxidicula*). The cited authors inferred the earliest late Oligocene age of the assemblage mainly from the absence of genus *Pyxilla*, whose extinction was considered to be near the early–late Oligocene boundary. However, Barron, who studied diatoms of the Morawan Formation, argued that this assemblage is not similar in composition to the late Oligocene assemblage described from the Bering Sea (Baldauf and Barron, 1987). The impoverished assemblage of the Morawan Formation bears some similarity with the assemblage of the Kamenka Formation from the Komandorskie Islands, which is assigned to Subzone "a" of the *Rhizosolenia oligocaenica* Zone. However, the former assemblage lacks *Rhizosolenia oligocaenica* Schrader, *Eurossia irregularis* (Greville) Sims, *Asteromphalus oligocenicus* Schrader et Fenner, *Asterolampra schmidtii* Hajos, *Navicula udintsevii* Schrader, and *Costopyxis trochlea* (Hanna) Strelnikova. This feature and the absence of *Rocela vigilans* Fenner allow the assemblage to be correlated with that of Subzone "b" of the lower Oligocene *Rhisozolenia oligocaenica* Zone (Fig. 2). Such a correlation is supported by presence of *Rhizosolenia antarctica* Fenner, whose first

occurrence in the high southern latitudes is confined to the lower part of the *Rhizosolenia oligocaenica* Zone (Baldauf and Barron, 1991; Harwood and Maruyama, 1992) and whose last occurrence is recorded in the lower part of the *Rocella vigilans* Zone (Fenner, 1984). However, the assemblage from the Morawan Formation lacks *Cavitatus jouseanus* (Sheshukova-Poretskaya) Williams, probably, because of its selective dissolution or paleoecological peculiarities of flora development.

In the Japanese Islands, the most complete sequence of the upper lower Oligocene through the lower Miocene is represented by the Tokiwa Formation in eastern Hokkaido (Akiba *et al.*, 1993). It exhibits the continuous zonal succession from the lower Oligocene *Rocella vigilans* Zone to the middle lower Miocene *Thalassiosira fraga* Zone. The diatom assemblages of this formation include all zonal species of the mentioned zones, except for *Thalassiosira praefraga* Gladenkov et Barron.

The Japanese researchers (Akiba and Yanagisawa unpublished data; Morita *et al.*, in press) reported about Oligocene and early Miocene diatoms from the Tokomuro, Tatsukobu, and Tsubetsu formations of eastern Hokkaido. The Tokomuro Formation contains diatom assemblages ranging from the *Rocella gelida* Zone to the lower part of the *Thalassiosira fraga* Zone, i.e., from the upper Oligocene to the middle lower Miocene. The Tatsukobu and overlying Tsubetsu formations yielded assemblages also correlative with flora found in the interval of the upper Oligocene *Rocella gelida* to the lower Miocene *Thalassiosira fraga* zones (all zonal species are present). The assemblage of the *Thalassiosira praefraga* Zone (the lower half of the lower Miocene) may be correlated with the diatom flora found by the mentioned researchers in the upper part of the Sankebetsu Formation of northeastern Hokkaido, though its index species are absent in this unit.

The diatom assemblages of the middle–late upper Miocene age are known from some localities of the eastern and southern parts of Honshu Island (Akiba, 1980, 1986; Yanagisawa, 1993; Akiba and Yanagisawa, unpublished data and personal communication). Deposits of the Hota Group in Boso Peninsula, eastern Honshu, enclose diatoms ranging in age from the lower part of the *Thalassiosira fraga* Zone to the middle part of the *Crucidenticula kanayae* Zone. Diatoms from the Yamanouchi, Kamenoo, and Taira formations; the Ikuchise Group; and the lower part of the Morozaki Group can be assigned to the *Crusidenticula sawamurae* Zone. The assemblage of the *Crucidenticula kanayae* Zone (the upper lower Miocene) characterizes the upper part of the Morozaki Group, the Shirado Group, the greater part of the Takaku Group, and the lower part of the Emi Formation (assemblages from higher horizons of the two latter subdivisions are correlated with flora of the *Denticulopsis praelauta* Zone).

CONCLUSION

This study demonstrated the possibility to use the diatom zonation (eight zones) for subdividing the Oligocene and lower Miocene deposits in different areas of the high–middle latitudes of the vast North Pacific region. The Oligocene and lower Miocene sections of the North Pacific and its framing structures are first correlated on this zonal basis. The recognized zones serve as a basis for identifying some phases in the diatom flora evolution and biogenic siliceous sedimentation in the region. In particular, the presented data suggest the intensification and wide distribution of biogenic siliceous sedimentation since the early Oligocene. This event coincided with the onset of global cooling, which resulted in the appearance of the cryosphere and psychrosphere.

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Magnetostratigraphy and Palynology of the Lower Pleistocene Loess–Soil Sequences in the Upper Don River Basin

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Abstract—A geomagnetic excursion is detected and studied in detail in the loess–soil sequence of the upper Don River basin. This event, termed as the Log Krasnyi excursion, happened about 600 ka ago. Characteristic features of this reversed polarity excursion consist in stepwise changes of polarity and lower intensity (by 3–4 times) of the geomagnetic field, as compared to this parameter of the stationary field. Palynological data show that the formation period of studied rocks included 7 warming and 9 cooling phases. The time range of the first half of the excursion corresponded to the period of cold climate, whereas its later stage consisted of two optimum phases. The commencement and cessation of the excursion are correlative with abrupt climatic changes. It is demonstrated that the Log Krasnyi excursion can be used to refine the Cromerian stratigraphy.

Key words: *magnetostratigraphy, paleomagnetism, excursion, chron, palynology, climatic optimum, extreme cooling, correlation.*

INTRODUCTION

Excursions of the geomagnetic field are very valuable in Quaternary stratigraphy. In addition to geological data, they serve as independent datum levels of a high resolution capability in subdivision and correlation of Quaternary sequences (Ryan, 1972; Pospelova *et al.*, 1976; Bucha *et al.*, 1977; Lazarenko *et al.*, 1980; Adamenko *et al.*, 1981; Zubakov *et al.*, 1982). However, the number of excursions, their position in the magnetochronological scale, characteristic features, and factors of origin are inadequately known (Fotiadi *et al.*, 1982; Braginsky, 1984; Gurarii *et al.*, 1986; Pospelova, 1990; Petrova *et al.*, 1992). Time ranges of the late Pleistocene and Holocene excursions have been studied better than the early Pleistocene interval (Kulikova *et al.*, 1979; Champion *et al.*, 1981; Pospelova *et al.*, 1986; Nami, 1995). Available data suggest that the stationary geomagnetic field was characteristic for a long period of the early Pleistocene, and excursions are confined only to the terminal interval of this stratigraphic epoch (Petrova *et al.*, 1992, and others). When known, the characteristic features of the early Pleistocene excursions and their correlation with climatic fluctuations may be very important for stratigraphy of the Cromerian period. From this viewpoint, the magnetic, paleomagnetic, and palynological studies performed for an indicative natural object seem to be of prime significance.

In this work, we used the aforementioned approach to study the loess–soil sequence in the upper Don River basin. Sequences of this type, as compared to other continental deposits, reveal the most complete record of paleoclimatic history of the Earth for the last 2 Ma (*Stratigrafiya*, 1982). At the same time, the loess–soil sequences are very suitable when we need to investigate the fine structure of geomagnetic records (Pospelova *et al.*, 1976; Tret'yak, 1983; Petrova *et al.*, 1992; Semenov, 1994; and others).

GEOLOGY OF THE REGION

The studied Pliocene–Pleistocene sequence of the upper Don River basin includes a complex of alluvial deposits of an intricate structure. These deposits are coeval to subaerial formations deposited outside of river valleys and consisting of alternating loess and fossil soil beds. The Don moraine is overlain here by the interglacial Muchkap deposits bearing remains of small mammals of the Tiraspol assemblage (Agadzhanyan, 1992), and this suggests that the region went through an early Pleistocene glaciation stage, presumably correlative with the oldest glacial epochs of Belarus and the Baltic regions. The regional stratigraphic scheme of the lower Pleistocene deposits in the upper Don River basin has been elaborated in detail, and almost all stratigraphic units of the East European scale, except the Okaian Horizon, have been distinguished in the

regional reference sections (*Resheniya*, 1986; Breslav *et al.*, 1992).

The lower Pleistocene deposits of the upper Don River basin are divided into six horizons (from the base upward):

(1) The Petropavlovka (*pp*) Horizon bearing remains of small mammals of the Tiraspol fauna, which includes the first representatives of the *Microtus* species;

(2) The Pokrovka (*p*) Horizon with characteristic periglacial spore–pollen spectra, cryophilic mollusks, and bones of small mammals of the Tiraspol fauna (the Pokrovka subassemblage);

(3) The Il'inka (*il*) Horizon of an intricate bed sequence spanning the interval between the Pokrovka and Don horizons and comprising up to three warm interglacial epochs; in subaerial facies, it corresponds to the loess–soil sequence including up to three beds of fossil soils, whereas in alluvial facies of the Don River basin, for instance, in the stratotype section of the Il'inka Formation located in the southeastern Voronezh region west of the town of Kalach, fossils from the horizon revealed three stages in evolution of the middle Tiraspol fauna of small mammals, which are younger than the Petropavlovka subassemblage (Krasnenkov *et al.*, 1992);

(4) The Don (*dns*) Horizon including morainal deposits of the Don glacier tongue together with fluvioglacial beds underlying and overlying them;

(5) The Muchkap Horizon consisting of interglacial alluvial deposits with the late Tiraspol microteriofauna (the Muchkap subassemblage), tortoise remains, and thermophilic mollusks;

(6) The Okaian Horizon of the lower Pleistocene deposits overlying the Muchkap Horizon.

In its details and microteriological characteristics, this scheme of subdivision of the lower Pleistocene deposits in the region is well correlative with the West European stratigraphic scheme accepted for the Cromerian Stage (Zagwijn, 1985).

To perform our paleomagnetic and palynological investigations in the upper Don River basin, we chose subaerial formations whose stratigraphic interval is defined by two datum levels. The upper level is the glacier tongue of the Don moraine, whose age is ≥450 to ≥490 ka according to thermoluminescence dates obtained by A.I. Shlyukov. The lower one corresponds to the Matuyama–Brunhes reversal detected earlier in the regional sequence by Semenov (1990) and additionally proved by our measurements.

SAMPLING PROCEDURE

We selected about 400 oriented samples for paleomagnetic and magnetic measurements from the most complete section of the lower Pleistocene deposits of the Log Krasnyi locality near the village of Uryv, in the Ostrogozhsk district of the Voronezh region (Fig. 1). Four prospect trenches I–IV partially supplementing one another (Fig. 1) were excavated in two gorges connected with the Izmaila River valley (a small tributary of the Don River). They exposed the Il'inka Horizon underlying the Don moraine and enclosing three beds of gray fossil soil. The soils are apparently unaltered and separated by loess beds. The third (lower) soil rests with a hiatus on a thin loam bed affected in part by pedogenesis. All these beds overlie either the red-brown soil or the light red alluvium of the Petropavlovka Horizon resting on the Cretaceous deposits. The basal alluvium of the Log Krasnyi section (Fig. 1) yielded the abundant and very diverse tooth remains of small mammals of the Tiraspol assemblage, including those of the oldest *Microtus* forms (Krasnenkov, personal communication, 1995).

In prospect trench I, which exposed the 13-m-thick sequence of three gray-brown soils with loess beds separating them and also the thin bed of loess-like loam and basal alluvium of the Petropavlovka Horizon, all underlying the Don moraine, we obtained oriented samples from 33 levels taking 2 to 5 rock cubes (144 in total) with 2-cm-long facets. Continuous sampling was performed for the second soil bed below the Don moraine, where we previously detected two levels with anomalous vectors of natural remanent magnetization. Prospect trench II is 250 m away from the first one and lower in the gorge valley, where, below the moraine, it exposes the 8-m-thick section with the second and third gray soils and loess interbeds. This sequence overlies the red-brown soil resting on the red and light colored Petropavlovka alluvium. The upper (submorainal) soil bed is presumably displaced by the moraine (Krasnenkov *et al.*, 1993). We collected samples 215 here, and each of 52 sampled levels was represented by 3 to 8 samples. Continuous sampling was applied to the second gray soil and to the upper part of the underlying loess bed. The small prospect trench III is located 3 m away from the second one and upstream in the gorge valley. We collected here 13 samples characterizing three levels in the basal interval of the second soil and loess bed underlying it. Trench II faces the last trench IV on the counter slope of the gorge, where the latter exposes the 6-m-thick section composed of red-brown soil and underlying alluvium, and we collected 23 samples from 5 levels of both beds. Samples for spore–pollen analysis (21 samples in total) were collected in trench II from the rock specimens simultaneously used to obtain the oriented samples for paleomagnetic measurements (Figs. 2, 7). Additional samples for palynological investigation were collected in the same manner from trench IV.

INVESTIGATION METHODS

Magnetic susceptibility K of the rocks was measured first *in situ* by using the equipment unit KT-5 and then with the laboratory susceptibility bridge KLY-2 in order to evaluate the magnetic anisotropy of the rocks

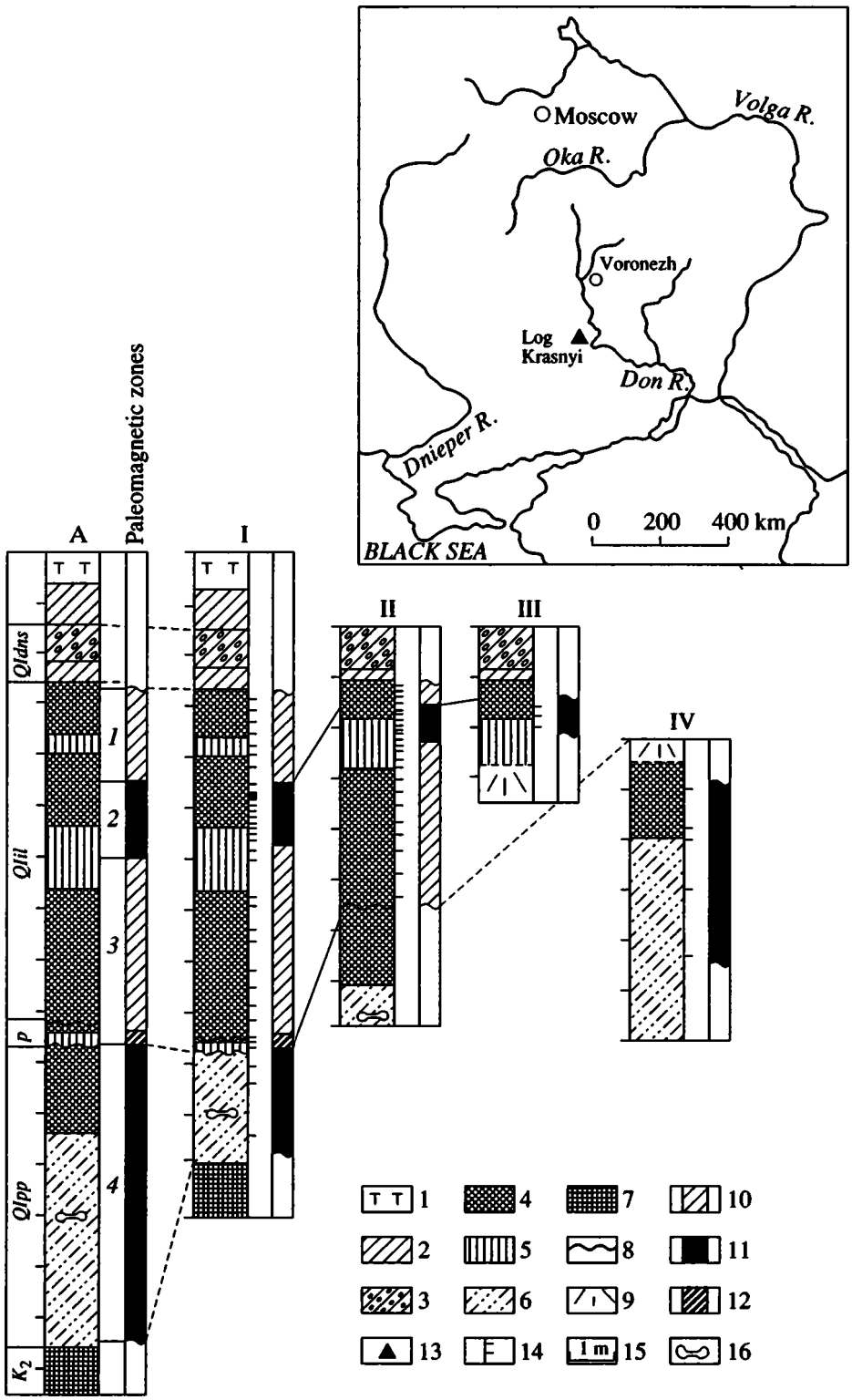

Fig. 1. Geographic position of the Log Krasnyi locality and magnetostratigraphic correlation of studied sections: (1) recent soil; (2) loam; (3) morainal loam; (4) fossil soil; (5) loess; (6) alluvium; (7) chalk; (8) hiatus; (9) talus; (10–12) paleomagnetic zones of normal (10), reversed (11), and transitional (12) polarities; (13) Log Krasnyi locality in the geographic scheme; (14) sampling level for paleomagnetic measurements; (15) vertical scale; (16) beds with fauna; (A) composite section compiled on the basis of individual sections I, II, III, and IV.

along axes *X*, *Y*, and *Z* of the cubic samples. *Jn* vectors of natural remanent magnetization (NRM) were measured with a JR-4 magnetometer after the sample screening for a period longer than a month. Directions of primary remanent magnetization were defined by using the thermal demagnetization procedure, when samples were heated stepwise up to 600°C in the laboratory demagnetizer, compensating the magnetic field to the level of 5γ, and afterward *Jn* components were analyzed.

The thermomagnetic method was used to analyze the composition of magnetic minerals. Blocking temperatures for magnetic minerals are identified in the *Jrs/K–(t)* plots depicting the dependence of remanent saturation magnetization (*Jrs*) on the temperature (*t*). *Jrs/K* ratio are used to estimate relative dimensions of magnetic mineral in the rocks.

To calculate relative intensity of the geomagnetic field at different stages of rock formation, we used the *Jn/Jrs* ratio together with values of factors $Qn = Jn/0.5K$ and *Q* at the temperature 400°C, when the *Jn* vector turned to be close to the initial direction.

In the technical treatment of samples for spore–pollen analysis, we followed the procedure of V.P. Grichuk usually applied to Quaternary sediments. Often, when spores and pollen grains of rock samples were surrounded by the mineral colloid, we used the more complex method of G.M. Levkovskaya. The highly colloidal and carbonate rocks were subjected to three-stage leaching in hydrochloric and hydrofluoric acids.

DATA OF MAGNETIC AND PALEOMAGNETIC INVESTIGATIONS

Rocks of the studied sections are heterogeneous in their magnetic properties. The scalar magnetic parameters *K*, *Jn*, and *Jrs* are obviously dependent on the rock lithology. In the red beds, they are three–five times higher than in the gray soils, loess interlayers, and light-colored alluvium. Magnetic susceptibility varies along the section from $110–250 \times 10^{-6}$ cgs, typical of the gray soils, loess, and light alluvium, to $500–1000 \times 10^{-6}$cgs, characteristic of the red-brown soil and red alluvial beds (Fig. 2). Anisotropy of magnetic susceptibility of the rocks is insignificant. Values of *Jn* and *Jrs* measured in the field of 143 000 A/m repeat the variation trend of the *K* parameter. The general range of *Jn* values is between 0.3 and 30*nT*, and only one sample of red-brown soil has this value equal to 62*nT*. Intervals of *Jrs* values are equal to 420–2700*nT* for the gray deposits and to 3000–5000*nT* for the red beds. Factor *Q* is also inconstant and usually varies from 0.05 to 1.5 in sections II and III and in the lower part of section I. The highest values of *Q* equal to 3–5 are obtained for samples of the light-colored alluvium. Magnetic parameters of the minimum magnitude are characteristic of the second gray soil below the Don moraine.

The studied rocks are different in the composition of their magnetic minerals. The *Jrs(t)* plots show that the fine-grained hematite is the main magnetic component of gray soils, loess beds, and light-colored alluvium (Figs. 3a, 3b), whereas this component in the red beds is represented by magnetite (Fig. 3c). The rocks are different not only in this aspect, but also in size and concentration of magnetic grains (Fig. 3d). The last figure demonstrates that data points in the *Jrs–K* plot tend to be scattered close to two straight lines. This means that all the samples can be divided into two groups: one with a lesser size of mineral grains (line "a") and another with greater mineral particles (line "b"). The first group includes samples of the light-colored alluvium, third gray soil, and loess bed overlying this soil. Coarse-grained samples from the second submorainal soil and, in part, from the underlying loess bed belong to the second group. Thus, the loess bed between the soils is divisible into two parts. At the same time, magnetic minerals are more concentrated in this loess bed and also in the soil below it, whereas at the second soil level (section II) they are less abundant. In other words, the concentration and grain size of magnetic minerals increase near the moraine base. This trend in mineral distribution is likely a consequence of moraine accumulation, because glacial waters appear to be able to wash out the fine grains of magnetic minerals from underlying deposits. The difference in composition of magnetic minerals is another problem. It may equally indicate the diverse origin of these minerals and/or the changeable provenance of sedimentary material.

After the stage of magnetic screening, *Jn* vectors display very scattered directions. Stereograms in Fig. 4a show *Jn* directions for different groups of samples. The maximum dispersion in this parameter is characteristic of the second submorainal soil and nearest levels of the underlying loess bed (Fig. 4a, II).

As one can see from the Zijderveld vector diagrams, which are plotted on the basis of data of stepwise thermal demagnetization, *Jn* vectors in samples from the normal magnetization zones have one (Fig. 5c) and sometimes two components. The first component ($T =$ 50–150°C) is incidental in its directions and represents the viscous remanent magnetization, which was not eliminated by magnetic screening of samples during the month period. Samples from the reversed magnetization zones usually reveal three components of the *Jn* vector (Figs. 5a, 5b, 5c). The first component, like in the samples with normal polarity, represents the viscous remanent magnetization, whereas the second one is a result of rock remagnetization in the present magnetic field and can be eliminated by heating up to 200–250°C, as follows from the component analysis. Differential vectors in the temperature interval of 150–250°C are compatible in their directions with the current geomagnetic field. The *Jn* directions abruptly change at the heating step above 200–250°C, and their concentration parameter increases (Fig. 4). Under the higher temperature, *Jn* directions change insignificantly, and lines

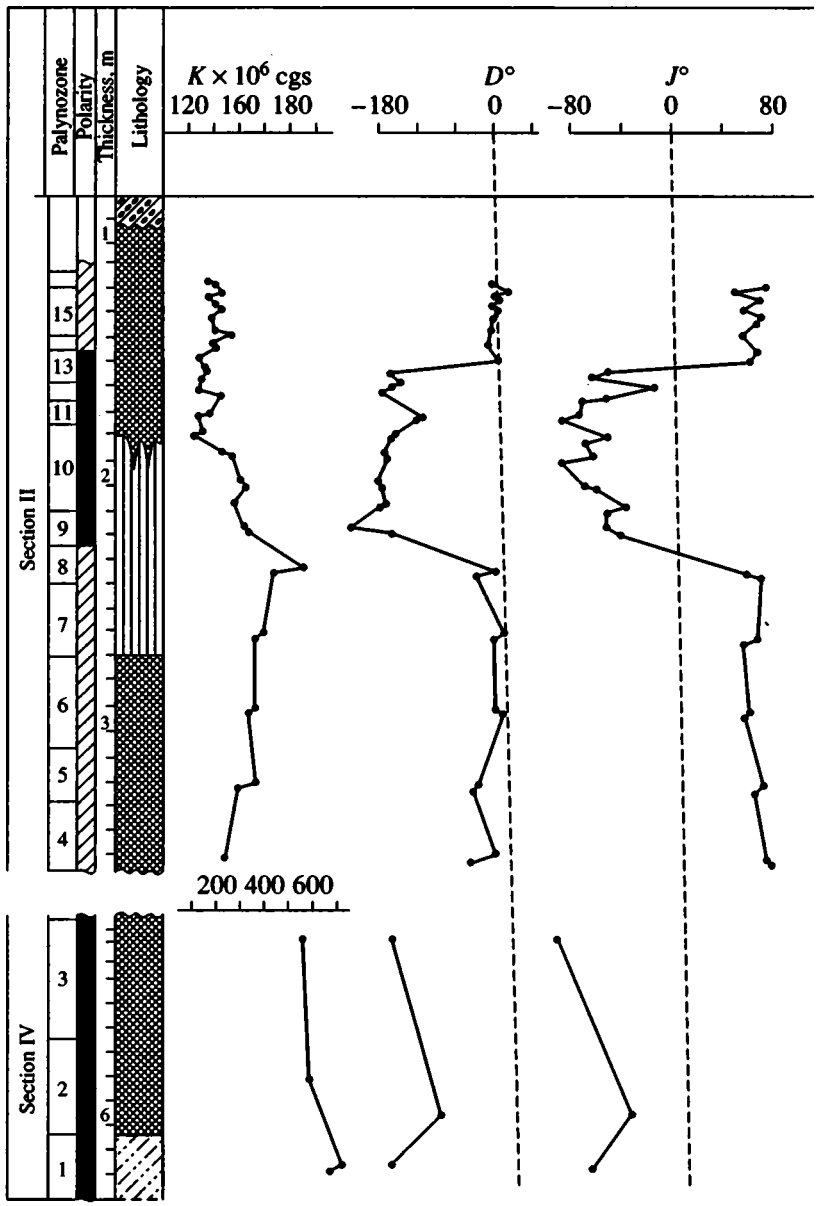

Fig. 2. Variations of paleomagnetic parameters of rocks versus paleomagnetic and pollen zones (symbols as in Fig. 1).

connecting projections of *Jn* vectors in the Zijderveld diagrams go toward the origin of the coordinates. Directions of characteristic magnetization, which are considered to be primary, are shown in Fig. 4b. One of the arguments enabling these directions to be interpreted as primary is the presence the same ferromagnetic (hematite) responsible for remanent magnetization of rocks in the two normal paleomagnetic zones and the reversed subzone.

In Section I, we distinguished four paleomagnetic zones (PMZ) different in thickness (Fig. 1). The first soil under the moraine along with the loess bed underlying this soil and an upper portion of the second soil compose the 2.1-m-thick PMZ-1 of normal polarity. The rest of the second soil and an upper part of loess bed below it are the rock units of the reversed PMZ-2, whose thickness is about 1 m. The 4-m-thick PMZ-3 of normal polarity begins inside the last loess interlayer and includes the third submorainal soil. Alluvium of the Petropavlovka Horizon (apparent thickness is 2 m) corresponds to PMZ-4 of the reversed polarity. A transitional zone 0.15 m thick is confined to the thin loess bed of the Pokrovka Horizon resting on the Petropavlovka alluvium and detected in three localities. In other sections of the upper Don River basin, Semenov (1990) also recognized a transitional zone between intervals of reversed and normal polarities above the Petropavlovka Horizon. In section II, the first submorainal soil and

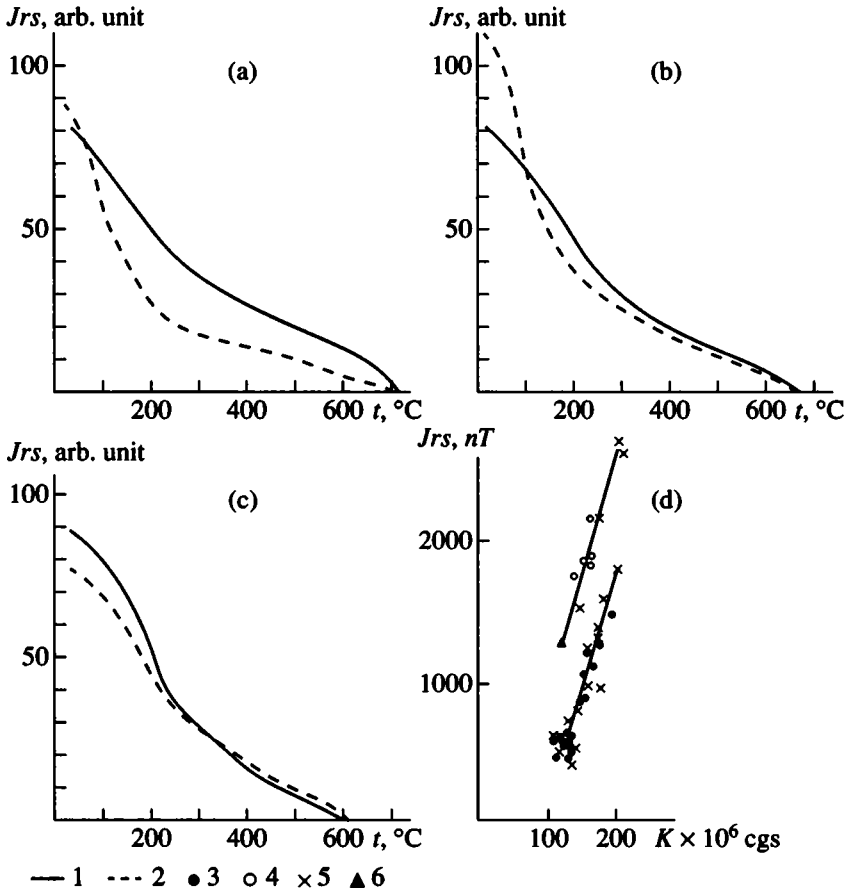

Fig. 3. Thermomagnetic curves (a, b, c) and *Jrs* variations versus magnetic susceptibility of rocks (d): (1) first heating run; (2) second heating run; (3) second gray soil in sections I and II; (4) third gray soil in Section II; (5) loess; (6) light-colored alluvium; thermomagnetic curves in plots A, B, and C are for the gray soil, loess, and red-brown soil, respectively.

underlying loess bed are missing. Here, PMZ-1 corresponding to the upper part of the second soil is only 0.5 m thick. The lower interval of this soil and the upper part of the underlying loess interlayer compose PMZ-2 of reversed polarity (0.7 m), like in Section I. Another part of the loess interlayer and the third submorainal soil again correspond to PMZ-3 of normal polarity, which is 3.3 m thick in this case. We also detected PMZ-2 of an inconsiderable thickness in Section III, where it is confined to lithological units correlative with their counterparts in sections I and II. The red-brown soil and alluvium of the Petropavlovka Horizon of Section IV represent the 4-m-thick PMZ-4 of the reversed polarity.

Paleomagnetic data for four sections are used to compile the composite paleomagnetic section (Fig. 1A). This section can be considered as consisting of two polarity zones different in thickness and separated by the transitional interval. The lower zone is about 6 m thick, invariably shows the reversed polarity, and corresponds to an upper interval of the Matuyama Chron. The Jaramillo Subchron is missing in this interval, because its level should correspond to beds with the Nogaiskaya fauna. The Matuyama–Brunhes boundary is recorded inside the loam bed of the Pokrovka Horizon underlying the base of the Il'inka Horizon with a hiatus. The upper paleomagnetic zone of normal polarity is 7 m thick and was formed during the first half-period of the Brunhes Chron. In the middle part of this zone, there is the reversed polarity subzone detected in three sections. The magnetization direction in this subzone is completely inverse relative to the normal polarity direction in adjacent intervals of the section. According to this feature, the reversed polarity subzone can be classified as a subchron of the Brunhes Chron, but at this stage of our investigation, we consider it as the Log Krasnyi excursion of the geomagnetic field, conventionally termed, after the geographic locality of observations. This excursion seems to be chronologically correlative with the Elunino-VII, Ureki-II, and Delta excursions; however, we cannot insist that they all are absolutely identical, because no detailed information about geomagnetic field fluctuations during the respective time periods is available. Nevertheless, the measured value of geomagnetic field intensity in the subzone is 3–4 times lower than that of the stationary field around it (Pospelova *et al.*, in press), and this is one of character-

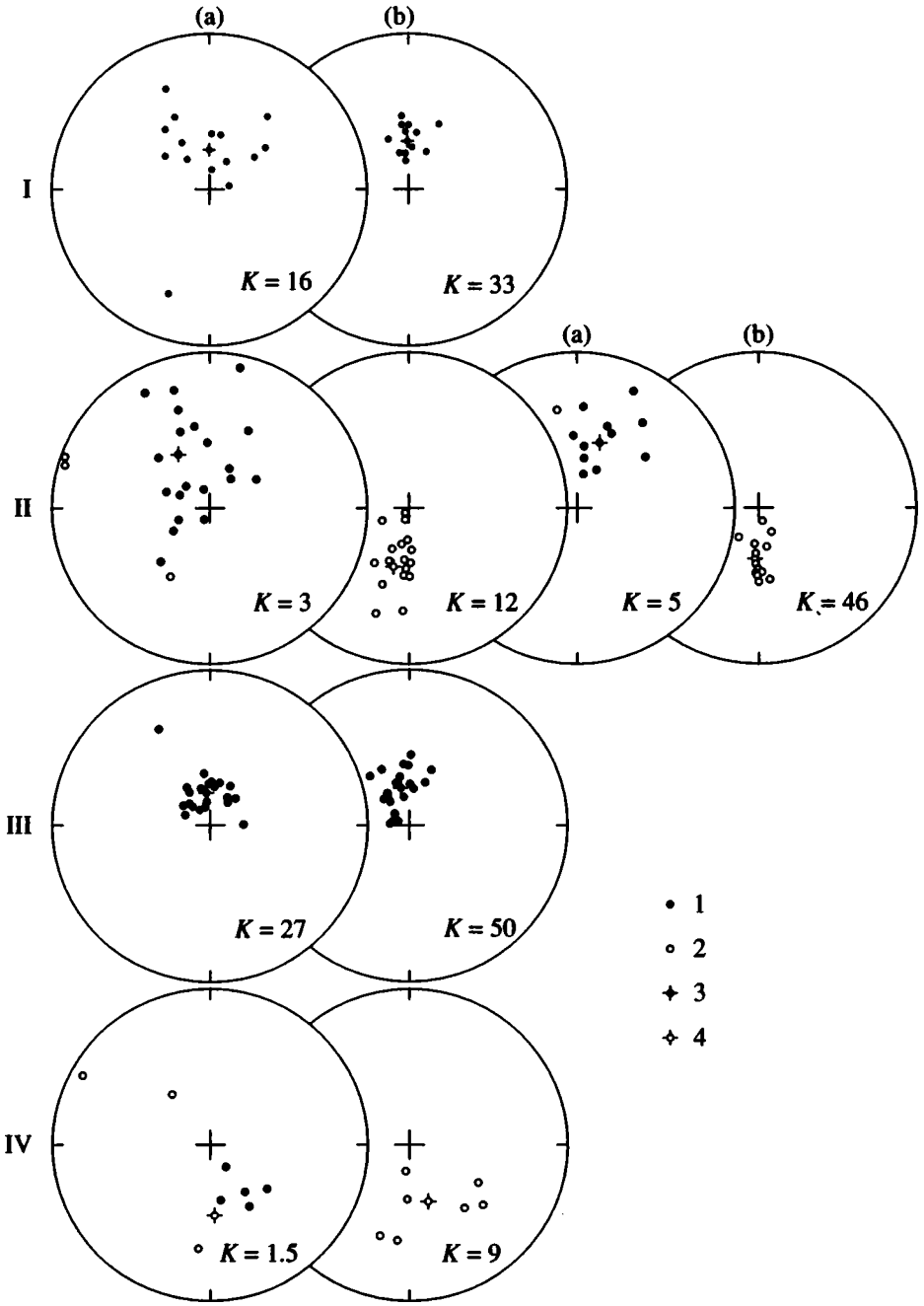

Fig. 4. Stereograms of *Jn* vectors before (a) and after (b) demagnetization: (1) projections on the lower hemisphere; (2) projections on the upper hemisphere; (3) mean direction of *Jn* vector in the lower hemisphere; (4) same in the upper hemisphere; I, PMZ-1; II, PMZ-2 in two parallel sections; III, PMZ-3; IV, PMZ-4.

istic features of geomagnetic excursions. Suggesting that the reversed subzone marks the terminal episode of the Matuyama Chron, we meet with difficulties concerning the interpretation of microteriological data. The Log Krasnyi excursion is recorded at the occurrence level of the middle Tiraspol fauna, and the early Tiraspol fauna is typical of the underlying deposits of the Petropavlovka Horizon. At the same time, the Jaramillo Subchron is confined to beds bearing the considerably older Nogaiskaya fauna, which represents an upper subassemblage of the complex Taman fauna. This subchron is also detected in deposits of the Ostrogozhsk Formation exposed in the Karatoyak section (20 km away from the Uryv locality) of the upper Don River basin, where these rocks have yielded a rich collection of the Taman microteriofauna, including its Nogaiskaya subassemblage (Iosifova *et al.*, 1992).

The excursion age can be approximately estimated on the basis of lithological, stratigraphic, and oxygen isotope data. Krasnenkov (1992) correlated the

regional stratigraphic scheme of Pleistocene deposits in the upper Don River basin with the general oxygen isotope scale of Shackleton and Opdyke (1977). As he has suggested, the Don Horizon presumably correlates with Stage 12 of the last scale, whereas the Il'inka Horizon spans the interval of stages 13–17, and the second submorainal soil of our sections corresponds to Stage 15. Consequently, the Log Krasnyi excursion may be as old as 580–620 ka.

Another approach to evaluate the age interval of the Log Krasnyi excursion includes the calculation of sedimentation rates. Deposits occurring between the Matuyama–Brunhes boundary and Don moraine are 7.5 m thick (Fig. 1) without regard for possible hiatuses, and consequently, for the incomplete character of the section. Since deposits in this interval were accumulated over 240 ka, the calculated sedimentation rate V_{av} was 0.0313 mm/year, on average. Judging from this value, the excursion commenced about 600 ka ago, and its duration cannot be longer than 20–30 k.y. The last value appears to be overestimated, because the real sedimentation rate should be higher than the calculated V_{av} value.

In contrast to other excursions, the geomagnetic field was constantly of the reversed polarity during the entire interval of the Log Krasnyi excursion. In this period, the virtual geomagnetic poles (VGP) wandered around the South Pole inside the latitudinal circles of 68° S or 55° S, as calculated on the basis of measurement data for Section I or for sections II and III, respectively. If we average all the data for three sections, then the calculated VGP position is at the latitude $\phi = 82°$ S and longitude $\lambda = 2°$ E, i.e., the geomagnetic axis in the excursion period was off by 8° from the Earth's rotational axis, which is quite similar to the present state. The angular deviation of VGP_{av} is calculated to be 12.9°.

The gradual transition from one polarity to another was not observed at either the initial or the terminal stages of the excursion, and this may reflect three possible factors: the stepwise origin of geomagnetic reversals; the short-term character of transitional processes unrecorded in the studied sections; or insignificant hiatuses in sediment accumulation. On the basis of the calculated sedimentation rate $V_{av} = 0.0313$ mm/year, the time ranges between the lower zone of normal polarity and commencement of the excursion in Section II, and also between the excursion cessation and upper zone of normal polarity, are estimated as equal to 3.5 and 1 k.y., respectively. Thus, we believe that the Log Krasnyi excursion is well substantiated by our data, and this event is dated back to approximately 600 ka ago.

PALYNOLOGICAL CHARACTERISTICS OF ROCKS

The spore–pollen analysis was performed for 24 rock samples from sections of the Log Krasnyi

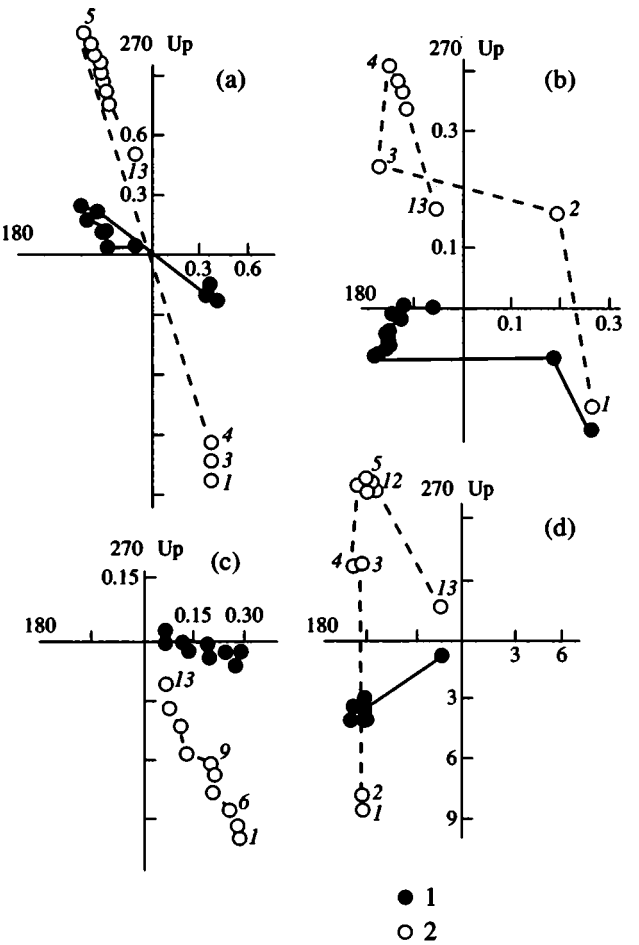

Fig. 5. *Jn* Zijderveld vector plots: (a) gray soil (Sample 22/1) of the reversed polarity zone; (b) loess (Sample 59/1) of the reversed polarity zone; (c) loess (Sample 7/4) of the normal polarity zone; (d) red-brown soil (Sample 42) of the reversed polarity zone; *Jn* projections (1) on the plane of X–Y and (2) on the plane of Y–Z. Numbers of data points indicate the heating temperature: (*1*) 20°C; (*2*) 50°C; (*3*) 100°C; (*4*) 150°C; (*5*) 200°C; (*6*) 250°C; (*9*) 400°C; (*12*) 550°C; (*13*) 570°C; *Jn* values along the axes are given in *nT*.

locality, including 3 samples from the reversed polarity zone of Section IV and 21 samples from Section II. The palynological data obtained are illustrated in Figs. 6 and 7.

Palynozone 1. The palynological spectrum of the red alluvium (Sample 24) shows a preponderance of arboreal taxa (82%) mainly represented by *Ulmus* pollen grains (64.4%). The latter coexist with rare specimens of *Quercus* sp., *Corylus* sp., and *Betula* sp. Pollen of *Juglans* sp. and *Carpinus* sp. is exotic. Pollen of aquatic plants is predominant in the herbaceous group. We identified water nymph (*Nymphaea* sp.), bladderwort (*Utricularia* sp.), soldier (*Stratiotes* sp.), and *Aldrovanda* sp. The well-preserved morphology of pollen grains suggests favorable conditions for their fossilization during the life period of the plants. This period

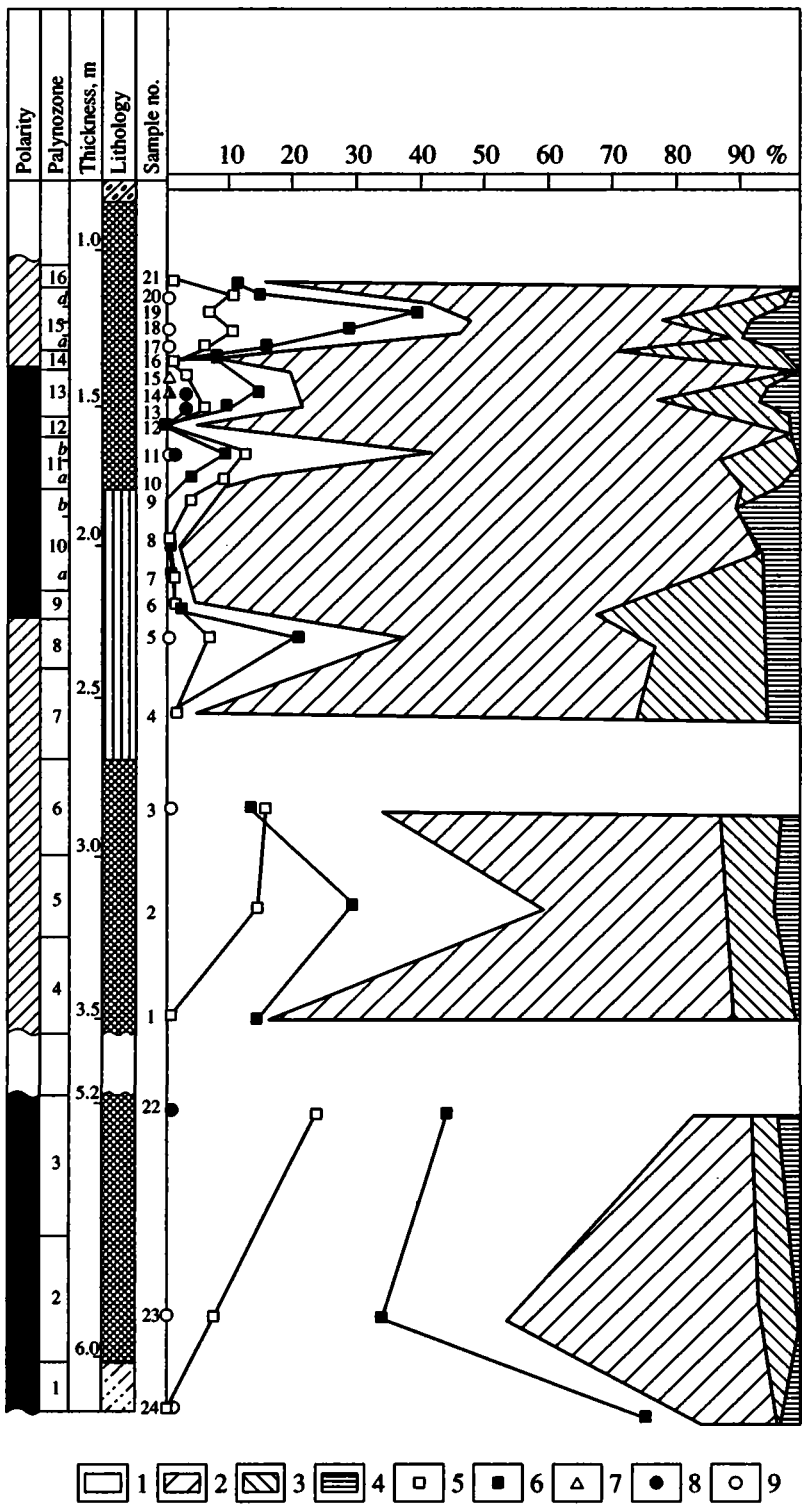

Fig. 6. Spore–pollen diagram for the composite section: (1) arboreal taxa; (2) mesophilic herbaceous–fruticose taxa; (3) herbaceous–fruticose xerophytes; (4) spores; (5) alder pollen; (6) total content of broad-leaved taxa; (7) spruce pollen; (8) pine pollen; (9) birch pollen (other symbols as in Fig. 1).

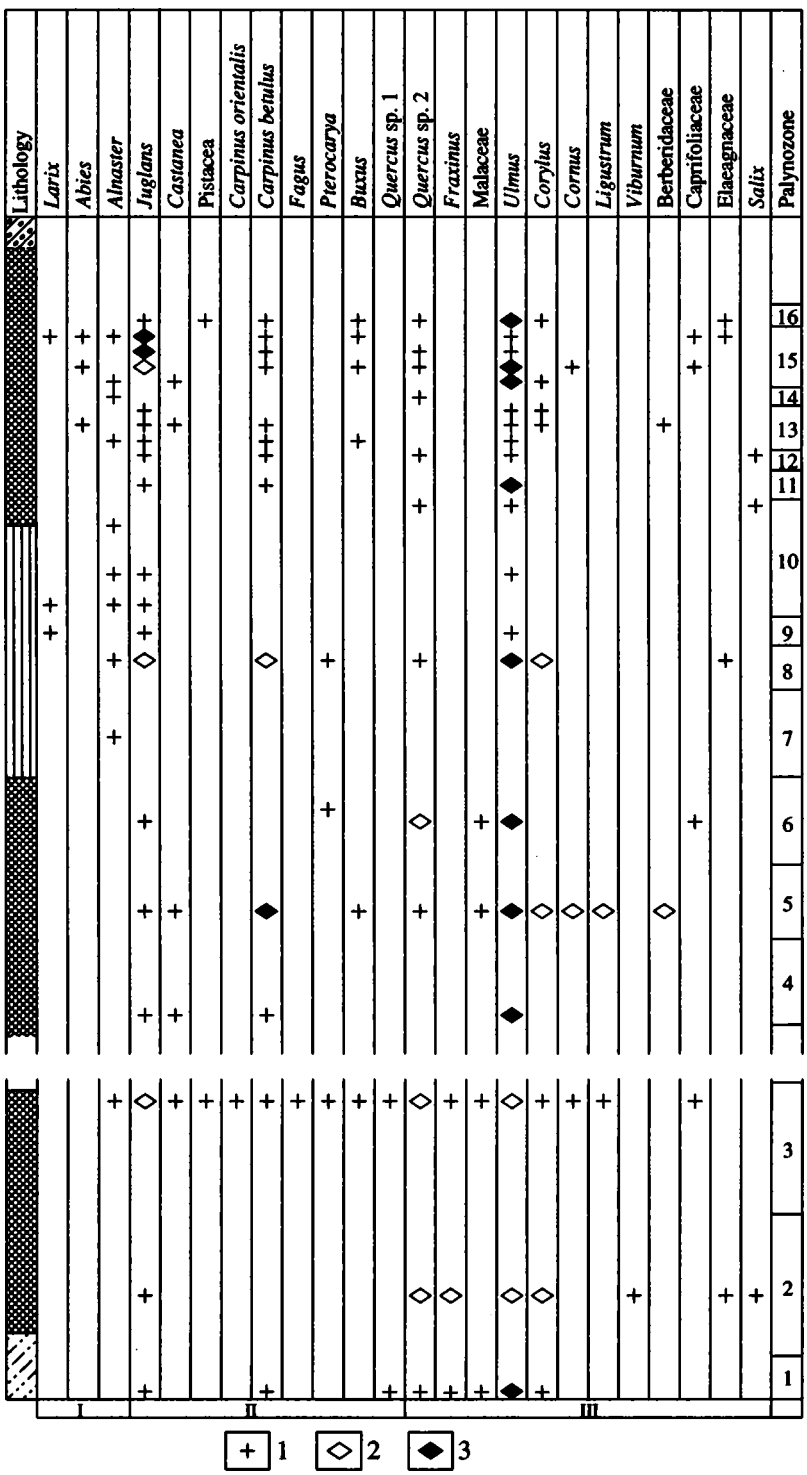

Fig. 7. Taxonomic composition of arboreal and fruticose pollen: I, microtherms missing in the present-day regional flora; II, thermophilic taxa missing in the present-day regional flora; III, other arboreal and fruticose taxa; (1) single grains; (2) abundant pollen; (3) dominant components of the broad-leaved arboreal group (other symbols as in Fig. 1).

corresponds to an optimum thermomer, when arboreal forest taxa were dominant in vegetative communities.

Palynozones 2 and 3. Palynological data for the red-brown soil (Fig. 6) show that the studied region with initial forest–steppe vegetation (Sample 23) was subsequently populated with pure forest communities (Sample 22). The content of arboreal pollen grows upward from 55 to 85%, and the abundance of herbaceous and fruticose taxa simultaneously decreases from 45 to 12%. At that time, preponderant broad-leaved polydominant forests were the specific feature of respective plant coenoses. Maple, oak, and ash trees, along with arbuscular Rosaceae (apple trees), Juglandaceae, and alder, dominated in regional forests during the first stage. Later, they coexisted with more hydrophilic arboreal plants. For instance, alder became more abundant, and singular pollen grains of beech, hornbeam, and chestnut appeared at this level. Nevertheless, elm, oak, and exotic Juglandaceae remained dominant among arboreal taxa. In general, dendroflora of this stage included very diverse exotic species (Fig. 7).

Pollen composition in the herbaceous–fruticose group also indicates a humidity increase in the region. This stimulated a preponderance of forbs, whose coenoses included *Ephedra, Artemisia,* Chenopodiaceae (Sample 23), and, later, *Carex* species (Sample 22). According to palynological data, the period of red-brown soil formation corresponded to the interglacial climatic optimum 3. In the classification of V.P. Grichuk, the first stage of flora evolution during this thermomer is an equivalent of the thermoxerophyte interstadial, and the second one corresponds to the thermohygrophyte climatic stage.

Palynozones 4–6. In general, the dominant pollen groups of palynospectra from the gray soil consist of either the mesophilic herbaceous–fruticose taxa (palynozones 4 and 6) or the arboreal specimens (optimum of Palynozone 5, Sample 2) associated with rather abundant pollen of forbs. We distinguished three subsequent periods of forest plant evolution named the *Ulmus, Alnus–Carpinus,* and *Alnus–Ulmus–Quercus* phases. Their exotic taxa are represented by *Buxus,* Juglandaceae, *Castanea,* and *Carpinus* species. The herbaceous group is rich in sedge pollen, especially at the end of the interval. Forbs are represented by Leguminosae, Plumbaginaceae, *Polemonium coeruleum,* Liliaceae, *Osmunda, Dipsacus, Armeria, Pachyloma,* Caryophyllaceae, Rosaceae, *Plantago,* Lamiaceae, and unidentified herbs (pollen of the latter is divided into 10 morphological types). In addition, we detected pollen of Gramineae and lacustrine–palustrine plants, such as avens (wet meadows, river banks, lake coasts, and bogs), water nymph, *Myriophyllium, Caltha,* and reed species accompanied by spores of green mosses. Pollen of *Ephedra,* Chenopodiaceae, and *Pachyloma* from samples of palynozones 4 and 5 suggests that regional landscapes included in the past isolated steppe and dry slope areas.

The considered flora resembles the present vegetation of the forest–steppe zone. At the same time, the forest coenoses of the optimum phase may imply that climatic conditions were warmer and more humid, i.e., less continental than at present.

Palynozones 7–10. These palynozones are distinguished by analysis of samples from the loess bed and lowermost soil layers above it (samples 4–10). The loess bed is heterogeneous in its characteristics and reveals certain levels with minimum abundance of arboreal pollen (palynozones 7, 9, 10) and one pollen horizon rather rich in this type of pollen (up to 23%), partially represented by broad-leaved taxa (Palynozone 8).

The pollen spectrum of Sample 5 (the last palynozone) suggests that principal open landscapes of the region coexisted with rather significant forest areas, because its composition is as follows: 38% arboreal taxa, 56% herbaceous–fruticose pollen, and 6% spores. The broad-leaved taxa (23%) were dominant among arboreal plants. *Ulmus,* the main taxon of the latter (27%), was more abundant than *Alnus, Corylus, Carpinus, Juglans graeca,* and apple trees, although they were all rather widespread. Cyperaceae and somewhat less abundant Chenpodiaceae were dominant in the herbaceous vegetation. Forbs were represented by Ericaceae, Cichoriaceae, Liliaceae, Lamiaceae, Droseraceae, Polemoniaceae, Fabaceae, Plumbaginaceae, and *Arctostapheyos.* Additional components are pollen of Gramineae and spores of green mosses, *Botrychium,* and *Sphagnum* species. Regional environments at this stage corresponded to those of the forest–steppe zone, but in contrast to present zones of this type, the past vegetation included exotic species of the Caucasus and of West European origin. This period can be classified as an interglacial optimum or warm interstadial.

Palynozones 7, 9, and 10 correspond to environmental stages almost free of forest vegetation. Their characteristics are rather similar, but not identical. Pollen is not abundant at these levels (5–15%) and includes such a microtherm as *Alnaster* sp. Samples 4 and 6 yielded larch pollen. At present, the nearest larch populations are either in the Carpathian Mountains, Karelia, Lapland, and Dvina–Pechora region, on the one hand, or in the Urals and Siberia, on the other. Cyperaceae were the main components of herbaceous plants (samples 4, 6, 9). These samples also recorded the maximum abundance of spores, and most plentiful among the latter are specimens of green mosses (these are especially numerous in Sample 4). *Botrychium* spores are rare. Flora of Palynozone 10, in contrast to the earlier forest-free stages (palynozones 7 and 9), was almost lacking in xerophytes in the herbaceous–fruticose plant communities. Consequently, we may assume a rather humid climatic phase for the accumulation period of the uppermost loess layers and basal soil strata.

Thus, the considered data imply a degradation stage of forest vegetation in the region. Landscapes of the respective time were apparently similar to those of the

present-day northern forest-tundra or tundra zones. Climatic conditions were cooler and more humid than at present.

Palynozone 11. Sample 11 yielded 44% broad-leaved arboreal pollen taxa and 56% palynomorphs of herbaceous and fruticose plants. The dominant arboreal forms are represented here mainly by elm and alder, whereas pollen of *Carpinus, Juglans,* and *Pine* is present as an accessory component. Forbs, especially *Plantago stepposa,* Leguminosae, and Liliaceae, are preponderant among herbaceous taxa. The first species now populating the solonetz and meadow steppe areas is of prime significance in this group. In addition, we detected pollen of Cyperaceae, Gramineae, and Chenopodiaceae. This palynospectrum is characteristic of forest–steppe areas and may indicate a phase of climatic optimum, when the elm forest areas expanded in the region.

Palynozone 12. The palynospectrum of Sample 12 consists exclusively of dwarf pollen of herbaceous and fruticose plants. Xerophytes are missing among them.

Palynozone 13. Palynological data for samples 13 and 14 suggest a somewhat greater abundance of forest areas in the region than at the previous stage, because the arboreal pollen content increased from 3 to 20%. Elm and hornbeam became dominant among the broad-leaved trees. The samples also yielded degraded carpolites of *Juglans*. In addition to broad-leaved forms, the forests in question included alder; pine and spruce appeared in their composition later. Herbaceous pollen spectra were initially (Sample 13) rich in Cyperaceae and also incorporated species of steppe (*Ephedra* sp. and Chenopodiaceae) and tundra vegetation (*Armeria* sp., *Dryas* sp., and Polemoniaceae). Afterwards (Sample 14), their dominant group consisted of forbs (Liliaceae, *Thalictrum, Ephedra,* Leguminosae, Chenopodiaceae, Caryophyllaceae, Polemoniaceae, Lamiaceae, and Pirolaceae). Pollen of Cyperaceae also became more abundant at this level and associated with other palustrine plants (*Stratiotes* and *Myriophyllium*). Both samples presumably reflect the thermoxerophyte and thermohygrophyte phases of the moderately warm interstadial.

Palynozone 14. Sample 16 representing this zone is identical in its palynological characteristics to Sample 12 bearing the dwarf pollen of herbaceous and fruticose plants.

Palynozone 15 (a, b, c). Data for Sample 17 mark a transition from the forest-free to the forest-rich stage of vegetation. In general, the thermomer of samples 17–20 corresponds to the period of broad-leaved forests with accessory alder and a mesophilic coenosis of forbs. The thermomer includes three subzones (Fig. 6) corresponding to successive evolutionary phases of flora and climate.

Phase 15a (Sample 17) represents the thermoxerophyte climatic period of the thermomer. Vegetation of this phase is marked by the peak abundance of herbaceous xerophytes (chiefly Chenopodiaceae) and also by elm predominance in the broad-leaved arboreal assemblage.

Phase 15b (samples 18 and 19) is an optimum of the thermomer, when forest areas were most widespread in the region. Dominant broad-leaved trees of this period were first represented by elm and then by *Juglans craeca,* whose pollen is of the exotic type.

Phase 15c (Sample 20) spans the terminal optimum interval of the thermomer. The general climate of the thermomer was warm and moderately humid and favorable for the appearance of interglacial forest–steppe landscapes.

Palynozone 16. Sample 21 revealed a pollen spectrum indicating the commencement of forest-free environments in the region. Pollen of broad-leaved arboreal taxa from this level seems to be redeposited.

Thus, paleogeographic environments during the formation period of our section were dissimilar. The obtained spore–pollen and botanical diagrams (Figs. 6, 7) display 16 palynozones that differ in abundance of arboreal pollen and other parameters. Palynozones 1, 3, 5, 8, 11, 13, and 15 correspond to optimum thermomers of various stratigraphic ranges, each marked by specific vegetation and climate. Regional landscapes of all these stages, except the earliest one, corresponded to the forest–steppe type, although they were not identical with respect to exotic and dominant taxa of forest vegetation. Palynozones 2, 4, 7, 9, 10, 12, and 14 mark the periods of climatic cooling. During these periods, forest areas degraded, and mesophilic forbs, together with Cyperaceae coenoses, were dominant in the region. The middle part of the loess bed underlying the second soil below the Don moraine (Palynozone 9) is the only level corresponding to more favorable time for xerophilic vegetation.

DISCUSSION

(I) Results of paleomagnetic measurements are correlated with palynological data in Figs. 2, 6, and 7. Inside the Matuyama Chron, comprising the section interval of red alluvium and red-brown soil, both with reversed parameters of magnetization, we distinguished two climatic optimums and one period of cooling. The third and second gray soils, together with the loess bed separating them, span the first half-period of the Brunhes Chron. In this interval, we detected five optimum and seven cooling stages. The third fossil soil with normal magnetic polarity was formed during an optimum in its middle part and during two cooling phases below and above this level. The lower interval of the loess bed overlying this soil is also in the normal polarity zone, and the basal portion of the interval was accumulated during a short cooling period, in contrast to the higher soil horizon deposited at the time of interglacial or interstadial optimum, i.e., during the climatic warming. The upper part of this loess bed is in the

reversed polarity zone and was deposited in the period of cooling like the basal soil horizon above the bed. In addition, the lower half of the second soil still situated in the reversed polarity zone reveals palynospectra of two short-term cooling and one warming episode. Finally, the upper horizons of this gray soil, which are already in the normal polarity zone, were first deposited under conditions of cooling and then during the subsequent interglacial optimum.

Hence, the first stage of geomagnetic excursion comprises a cooling period, which is longest and most extreme among those distinguished in the section. Its second stage spans two brief episodes of moderately warm interstadial or, let us say "interphasial," type separated by a short-term cooling phase.

The performed analysis of published and our own data (Petrova *et al.*, 1992) concerning the correlation between geomagnetic excursions and climatic fluctuations of the Brunhes Chron ended in controversial results. Geomagnetic excursions detectable in marine or oceanic sedimentary sections of the middle–upper Pleistocene are correlative with periods of climatic warming. Rampino (1979), who attempted to interrelate the excursions with secular changes in the eccentricity of the Earth's orbit, and consequently with climatic fluctuations, arrived at the same conclusion. According to his synthesis, all the excursions, including the one that happened about 600 ka ago, are coeval with eccentricity maxima of the orbit, i.e., with periods of global climatic warming. In continental sections, however, geomagnetic excursions are recorded in sediments deposited in both the cool and warm periods of sedimentogenesis. Climatic environments in the last case are depicted on the basis of rock lithology and palynological data without regard to their correlation with paleomagnetic episodes. For instance, the magnetic excursion Ureki-II is recorded at the base of the Paleoeuxinian beds of the Black Sea region (Zubakov and Kochegura, 1976) and correlated with the oxygen-isotope excursion of Stage 16 (Zubakov *et al.*, 1983). Its level was initially dated back to 390 ± 50 ka, then to 490 ± 50 ka, and finally to 620 ka, and the last value is close to the estimated age of the Log Krasnyi excursion. Nevertheless, there is no other locality, except that of the Log Krasnyi, where an excursion spans the climatic phases of cooling and warming.

On the basis of our data, we believe that possible climatic fluctuations during an excursion are less substantial than the climatic conditions immediately before and after this event. The commencement of our excursion is clearly coeval with the phase of significant cooling at the time of the oxygen-isotope excursion of Stage 16 which happened after the interglacial or warm interstadial phase, whereas the terminal cooling at the upper limit of the excursion is replaced by another interglacial stage of warm and moderately humid climate.

These significant climatic fluctuations at both ends of the excursion interval suggest that geomagnetic and climatic events actually could be interrelated. Our conclusion based on experimental results is in good agreement with the theoretical concept of Rampino (1979), who argued that swift inversions of the geomagnetic field (excursions) and climatic oscillations are the result of common astronomic factors. Nevertheless, this conclusion needs a more comprehensive substantiation.

(II) The established climatological and geochronological parameters allow the Log Krasnyi excursion to be used in correlation practice and subdivision of Cromerian deposits. In particular, stratigraphic, paleomagnetic, and palynological data on the lower Pleistocene sequences of the upper Don River basin provide important information when compared with similar data on Cromerian deposits in northwestern Europe (Zagwijn, 1985).

On the basis of palynological data, Zagwijn divided the Cromerian sequence of northern Europe into seven glacial and interglacial climatic stages (Fig. 8a) spanning together the stratigraphic interval from the Petropavlovka to the Muchkap Horizon in our region. The established thermoclimatic curve of the Cromerian shows glacial stages *A*, *B*, and *C* (one with a double peak), which are separated by interglacial stages I, II, III, and IV. The Interglacial IV seems to correspond to the Muchkap Horizon, whereas the period of significant cooling before it (glacial stage *C*) is a possible equivalent of the Don Horizon.

Below the Glacial *C*, Zagwijn distinguished two climatic optimums (interglacials II and III) and an extremum of cooling (Glacial *B*) with a subordinate optimum phase. Our palynological data for the Log Krasnyi section (Fig. 8b) allow us to conclude that the second submorainal soil was accumulated during the optimum period incorporating two minor cooling phases. This climatic optimum is presumably correlative with the optimum phase of the Glacial *B*. The upper part of the loam bed between the second and third gray soils marks a period of extreme cooling and may correspond to the first peak of the Glacial *B*. The lower part of this bed and third submorainal soil were deposited during the Interglacial II, which also incorporates two subordinate cooling phases. Accordingly, the first submorainal soil appears to be the stratigraphic equivalent of the Interglacial III. The general character of climatic oscillations revealed in the Log Krasnyi section on the basis of palynological data is more complex than in the scheme of Zagwijn (Fig. 8). The red-brown soil of the Petropavlovka Horizon, which belongs to the reversed polarity zone, bears palynoflora of the thermoxerophyte and thermohygrophyte interglacial phases and seems to be correlative with the Interglacial *I*. If this correlation is correct, then there is a considerable hiatus between the third gray soil and the Pokrovka Horizon, whose deposits were accumulated during the glacial stage *A*.

The Matuyama–Brunhes boundary is the only magnetostratigraphic reference level in the Cromerian deposits, and their upper beds corresponding to the

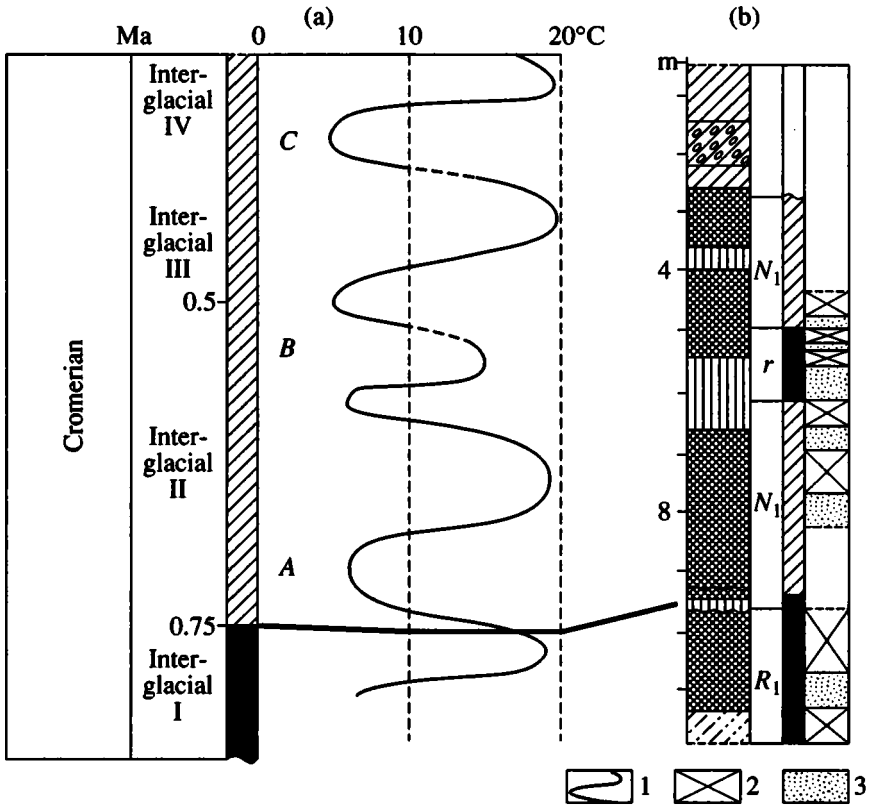

Fig. 8. Stratigraphic, paleomagnetic, and palynological correlation between (a) the lower Pleistocene loess–soil sequence in the upper Don River basin and (b) Cromerian deposits in northern Europe: (1) thermoclimatic curve after Zagwijn (1985); (2) optimum; (3) cooling extremum (other symbols as in Fig. 1).

Brunhes Chron are lacking records of geomagnetic excursions. At the same time, the Log Krasnyi section provided an opportunity to define the age and climatic characteristics of such an excursion and to establish its position in the Cromerian interval. The excursion is correlative with the first cooling phase of the Glacial *B* and, partially, with the initial time of the minor optimum of this glacial stage. Their total interval is older than 0.5 Ma and younger than 0.73 Ma (Fig. 8), and, according to our evaluation, the excursion is approximately 600 ka old.

Thus, the established excursion represents an additional geomagnetic reference level useful for subdivision and correlation of Cromerian deposits.

(III) Some time ago (Pospelova *et al.*, 1994), we reported that scalar magnetic parameters of rocks are interrelated with climatic oscillations that occurred during the formation period of Quaternary sediments in the Pogrebya section (Moldova) and Treugol'naya Cave. The established correlation was characteristic of both the high-amplitude climatic oscillations and the subordinate climatic phases. In both cases, magnetite grains of the pseudomonodomain type were similar in composition and structure throughout the studied sections and represented principal rock components responsible for magnetic properties of all the beds. However, we failed to trace the same trend in the Log Krasnyi section (Fig. 2), because the homogeneous composition and domain structure of magnetic grains throughout the section is the necessary condition for revealing a correlation between the scalar magnetic parameters and palynozones. Unfortunately, the studied section is heterogeneous with respect to composition and size of magnetic minerals (Fig. 3).

CONCLUSION

(1) We detected and studied in detail the Log Krasnyi excursion, which originated about 600 ka ago. This is the first geomagnetic excursion recognized in the lower Pleistocene deposits. Its characteristic features are as follows: the constantly reversed polarity in the excursion interval, the stepwise transition from normal to reversed polarity and vice versa, a lower intensity of the geomagnetic field than that of the stationary field around the excursion, and a value of VGP deviation off the South Pole similar to the present one.

(2) Palynological data on rock samples used for paleomagnetic measurements indicate the variable paleogeographic environments of the rock formation. We distinguished 7 optimum and 9 cooling stages. The initial half-period of the excursion corresponded to the

cold climatic stage, the second half of which included two subordinate optimum phases. The commencement and cessation of the excursion are at the levels of abrupt transition from the interglacial stage to the extreme cooling phase and vice versa. This may suggest that climatic oscillations and geomagnetic excursion are interrelated.

(3) Comparison of climatic oscillations in the Log Krasnyi section with those depicted by Zagwijn (1985) in Cromerian deposits of northern Europe shows that the excursion interval spans the first stage of the Glacial *B* and, in part, the subsequent phase of warming. The excursion in question is correlative in time with oxygen-isotope excursion 16 and the lower part of excursion 15. The established chronostratigraphic position of the Log Krasnyi excursion offers an opportunity to seek this event in other localities and to use it in practice of correlation and subdivision of Pleistocene deposits.

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The Gothenburg Magnetic Excursion in the Black Sea Sediments

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Abstract—The late Pleistocene–Holocene deposits are studied in 12 piston-core sections sampled at the continental slope and in the deep part of the Black Sea within the Adler–Tuapse test area. In four of these sections, the Neoeuxinian beds revealed paleomagnetic anomalies confined to the base of the hydrotroilite member[1] and presumably induced by the Gothenburg excursion. Correlation with available data on the western region of the Black Sea, where the Gothenburg excursion was first identified, substantiate synchronism of hydrotroilite deposits in eastern and western areas of the sea and elucidate principal trends in the paleogeographic evolution of the Black Sea in the late Pleistocene–Holocene time.

Key words: *Black Sea, Pleistocene, paleogeography, paleomagnetism, Gothenburg excursion.*

STATE OF THE ART AND PURPOSE OF THE STUDY

Excursions of the geomagnetic field have been discussed in a series of publications, but many aspects of these events remain still ambiguous. For instance, it is disputable even now whether the excursions are global or regional. In addition to its significance in the theory and evolution of the geomagnetic field, an answer to this question is also important in the pragmatic aspect, because we should know if these excursions are applicable in interregional (global) stratigraphic and paleogeographic correlation, or their correlation potential is limited. Many researchers, especially those who are involved in generalization of available data on excursions, consider them as global on the basis of very serious arguments.

Pospelova (1989), who completed the statistical processing of published data on the period of 20–6 ka that were obtained for 105 sections studied in various sites of the Earth, detected one prominent peak corresponding to the interval of 11–13.5 ka on the histogram and concluded that the Gothenburg excursion was a real and global paleomagnetic event. As known, Mörner (1977), who was the first to discover this excursion, defined its age interval as corresponding to 13 750–12 350 years ago (the oldest Dryas). According to this investigator, the geomagnetic field fluctuations were irregular during the excursion, and there was only one short-term inversion (flip) for a period of about 200 years.

At the same time, some studies on this issue failed to reveal the Gothenburg excursion in the indicated time span. In particular, Zagnii and Bakhmutova (1986), who investigated the varved clay sections in Karelia and the Leningrad region, reported that "the Gothenburg event of reversed polarity in the interval from 10 000 to 16 000 years ago is missing here." This is remarkable, because the studied sections are located not far away from but close to the latitudes of the stratotype sections in Sweden, where the Gothenburg excursion was originally distinguished. Banerjee *et al.* (1979) were also unable to detect this excursion in the Minnesota Lake sediments of the same age.

From this viewpoint, it would be purposeful to study particular objects, e.g., the relatively continuous sections accumulated under sufficiently high sedimentation rates, since they are most suitable for the purpose of tracing such a short geomagnetic event. One of such objects is the Late Cenozoic section of bottom sediments in the Black Sea, because it can be easily sampled by piston coring and is known to span the interval of the Gothenburg excursion. The presence of this excursion in the Black Sea sediments was first suggested by Tret'yak *et al.* (1989); however, stratigraphic intervals of normal and reverse magnetization in the rocks, the latter assumed to be corresponding to the Gothenburg excursion, were not clearly outlined at that time, and an adequate identification of this geomagnetic event remained to be better substantiated. Later, Bol'shakov and Nikolaev (1993) recognized the Gothenburg excursion in the Neoeuxinian deposits of the western halistasis of the Black Sea. We want to stress that Bol'shakov and Nikolaev performed paleomagnetic measurements in carefully selected core sections

† Deceased.

[1] Under this term, we consider here the black deposits containing not less than 0.08% hydrotroilite.

that were well stratified and lacked possible marks of turbidity sedimentation. Accordingly, their identification of the Gothenburg excursion seems to be properly substantiated from the geological viewpoint at least in sections 11 and 1.

Stratigraphic subdivision and correlation of the late Pleistocene–Holocene deposits of the Black Sea can be found based on different criteria. In shallow areas rich in molluscan fauna, specialists use the very detailed scale elaborated by L.A. Nevesskaya and E.N. Nevesskii at the beginning of the 1960s, when sediments were divided into the Neoeuxinian, Bugazskii, Vityazevskii, Kalamitskii, and Dzhemete beds. The deep-water deposits are free of fossils due to hydrosulfuric pollution; A.D. Arkhangel'skii and N.M. Strakhov suggested at the end of the 1930s to divide them into the Neoeuxinian, Paleochernomorian, and Recent beds on the basis of lithological criteria. Later, Shcherbakov *et al.*, (1978) attempted to correlate both schemes in order to obtain a more detailed scale for the deep-water sequences divisible on the basis of sediment lithology. However, lithological criteria used in the stratigraphic subdivision of deep-water sediments in the western part of the Black Sea are inadequately distinct in its eastern areas. For instance, the Kalamitskii sapropel beds grade in this direction into sapropel-like deposits and become externally similar to the Dzhemete beds. Sedimentation in the latter areas is further complicated by the turbidity, avalanche, and other gravity-induced processes that are more frequent here than in the west. In addition, terrigenous sedimentary components are more widespread in the east because of the proximity of the Caucasus Mountains.

At present, there are more than two hundred published age determinations for boundaries separating the recognized stratigraphic units. However, the results are controversial; in this work, we accept data of Nikolaev *et al.* (1980), who inferred the following age values for upper boundaries of the units: 9 ka for the Neoeuxinian Horizon, 7–8 ka for the Vityazevskii–Bugazskii beds, and 3.5 ka for the Dzhemete beds. According to Shcherbakov *et al.* (1978), the base of the Neoeuxinian Horizon is approximately as old as 17 ka.

Sediments accumulated without the influence of gravity processes are extremely rare in the Black Sea. It is a classical example of a sea basin with abundant and diverse gravity and avalanche sedimentation. Many sedimentary sequences here consist of turbidites and slope facies. They display frequent erosion marks and hiatuses, some of which may correspond to a whole horizon, and these features greatly complicate subdivision and correlation of deep-water sediments. Moreover, the radiological dating of the Neoeuxinian strata is almost impossible, because disequilibrium methods are inapplicable due to initially discordant isotopic ratios of the U–Th series in sediments, and radiocarbon dating is misleading, since the rocks are depleted in organic matter and always contain significant but variable proportions of old terrigenous carbonates.

Thus, the paleomagnetic methods of investigating the magnetic excursions seem to be most promising for solving problems of stratigraphy of deep-water deposits in the Black Sea. Nevertheless, one should be very careful when identifying the fine peculiarities of geomagnetic trends, such as excursions, in deep-water sediments of this sea. We believe that slope sedimentation and contour currents may distort the paleomagnetic records and be responsible for anomalous magnetization vectors not related in origin to possible changes in the geomagnetic field. Taking this into account, we decided to study not only special sections, which are free of specific features of the slope sedimentation, but also ordinary core sequences obtained in the course of geological survey.

The main purpose of our study was to answer the following questions:

(1) Is it possible to detect paleomagnetic anomalies reflecting geomagnetic fluctuations in the slope sediments deposited under the influence of relevant factors?

(2) How widespread is the Gothenburg excursion in the Black Sea sediments?

(3) To what extent is it possible to correlate the paleogeographic evolution of eastern and western areas of the Black Sea using the Gothenburg excursion as the datum level, if we are able to recognize it?

MATERIALS AND INVESTIGATION PROCEDURE

The late Pleistocene–Holocene deposits studied here were sampled by piston coring from the continental slope and deep bottom of the Black Sea in the Bol'shoi Sochi (Adler–Tuapse) test area during the expedition of r/v *Yuzhmorgeologiya* in 1991. The detailed study of magnetic properties and paleomagnetic patterns was performed for 12 sampling sites shown in Fig. 1. As noted above, we investigated ordinary core sequences obtained during the geological survey. In all the sites, the cored upper Pleistocene–Holocene deposits suitable for detecting the Gothenburg excursion are rather thick (up to 7 m) and sometimes affected by different factors of the slope sedimentation. Samples as large as $2 \times 2 \times 2$ cm were loaded into polystyrol containers resistant against heating up to $t = 95°C$. The samples were subjected to thermomagnetic cleaning for 24–32 hours under a Permalloy screen and $t = 90°C$. As known (Faustov *et al.*, 1986), this procedure is sufficient to destroy the viscous magnetization induced during the last 15–20 ka. The magnetic anomalies, which are considered as related to the Gothenburg excursion, were detected in four sections, and respective data are shown in Figs. 2–5.

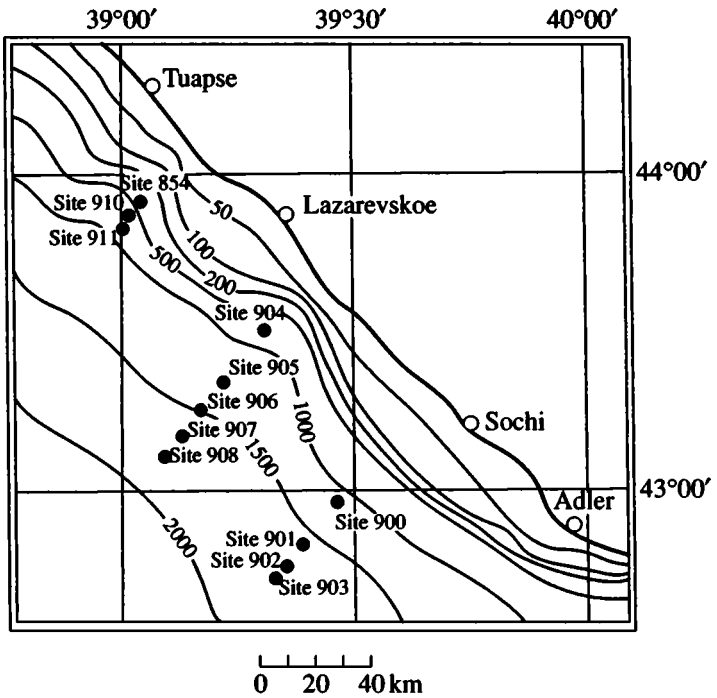

Fig. 1. Geographic scheme of the test area and localities of studied core sections.

DISCUSSION

Unfortunately, the influence of slope sedimentation and other factors on orientation of remanent magnetization vectors in the deep-water sediments is poorly studied. Rees *et al.* (1968) and Hailwood (1978) argued that, in the La Jolla submarine canyon and Bristol Channel, the orientation of remanent magnetization vectors could be changed under the action of contour currents. In addition, Faustov *et al.*, (1987) showed that magnetization vectors in some continental slope deposits are considerably dependent on the slope orientation and reflect either the influence of deflection currents or the result of the diluvial slope wash. From these viewpoints, it is not inconceivable that slope processes also affect the orientation of I_n vectors (mainly the magnetic declination) in the bottom sediments. As is known (Voskresenskii, 1971), the elongated particles (when they are magnetic, their elongation and magnetic moment coincide) tend to orient their longer axes in sediments parallel to the slope dip. Therefore, we may easily assume that paleomagnetic records in turbidites cannot be substantially distorted, since sedimentary particles precipitate in this case under quiet hydrodynamic conditions after the avalanche drop of the turbidity current. This is well evident from the common granulometric differentiation of turbidites. At the same time, if it is rightful to attribute terrestrial phenomena to the submarine slope sedimentation, the deflection currents at the ground depending on the slope orientation can change the I_n vector direction, first of all, the magnetic declination. These phenomena are characteristic even of gentle slopes (~3°–5°), and the angle between the slope dip and I_n can be variable and as large as one can imagine. Precisely this fact suggests that paleomagnetic interpretations of magnetic anomalies, especially when they depict only changes in I_n parameters, cannot be unambiguous and are related only with geomagnetic field reversals.

The results obtained for Site 910 and shown in Fig. 2 apparently illustrate the influence of slope sedimentation factors on the declination of I_n vectors. Naturally, we cannot determine the orientation of core sections relative to the magnetic meridian, when they are sampled by piston coring of bottom sediments. However, this unknown parameter is common for all the core samples, and we can detect changes in the magnetic declination relative to the conventional meridian throughout the cored interval. In the case when the uppermost beds of the core section are undisturbed, it is possible to correlate their declination vectors with the current parameters of the geomagnetic field and to recalculate the declination of I_n vectors in all core samples as fixed in the present coordinate system. We applied this procedure of data processing for the core section obtained at Site 910 (Fig. 2), where recent deposits were sampled in the least disturbed state, as compared to other sites. As one can see from Fig. 2, magnetic parameters of most samples are inconsistent with the present geomagnetic field. We do not know the precise slope orientation at Site 910, but, in general, it is of the southwestern direction close to 225°, as one can see from Fig. 1. Declinations of I_n vectors are generally oriented toward the west and northwest; i.e., they are in the intermediate position between the present

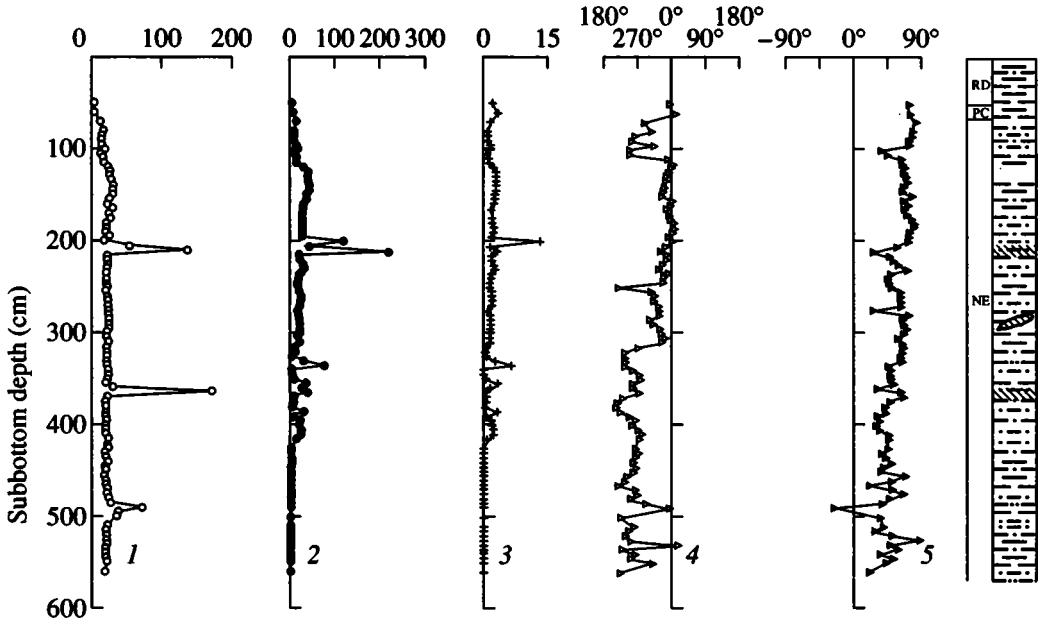

Fig. 2. Paleomagnetic characteristics of deposits recovered at Site 910 (see Fig. 5 for explanation).

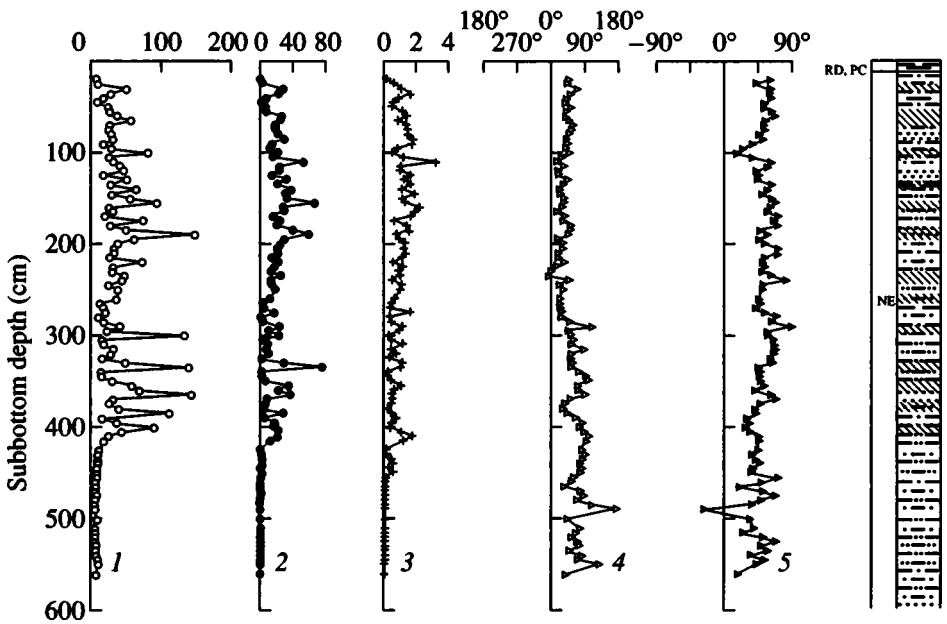

Fig. 3. Paleomagnetic characteristics of deposits recovered at Site 903 (see Fig. 5 for explanation).

direction of the geomagnetic field and the general orientation of the slope. This may imply that I_n vectors in deposits at Site 910 are induced under the mutual influence of the current geomagnetic field and slope sedimentation factors, as was once suggested for the slope loams (Faustov *et al.*, 1987).

These speculations show that it would be incorrect to correlate all anomalous magnetic parameters of sediments with excursions of the geomagnetic field.

In our analysis of paleomagnetic data on the studied sections, we suggest a magnetic excursion only in the case when changes in declination D are combined with changes in the inclination angle up to the inversion of its sign. We also estimated the lithological features of sediments, and when they appeared to be related to gravity sedimentation, the revealed paleomagnetic anomalies were not unambiguously attributed to the geomagnetic excursions. In contrast to overland environments, where beds with magnetic anomalies can be

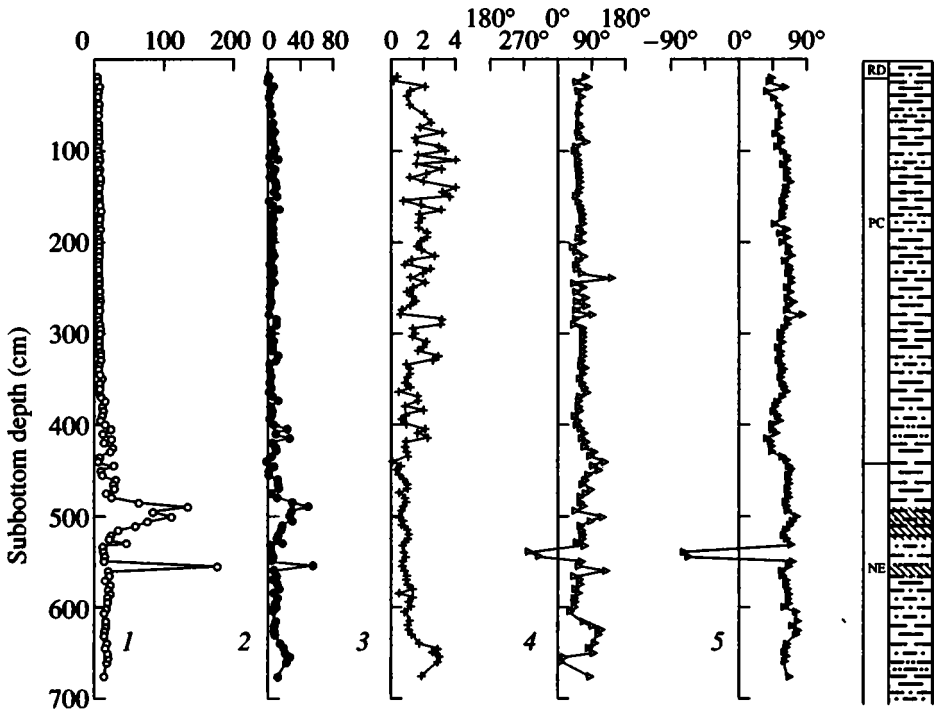

Fig. 4. Paleomagnetic characteristics of deposits recovered at Site 908 (see Fig. 5 for explanation).

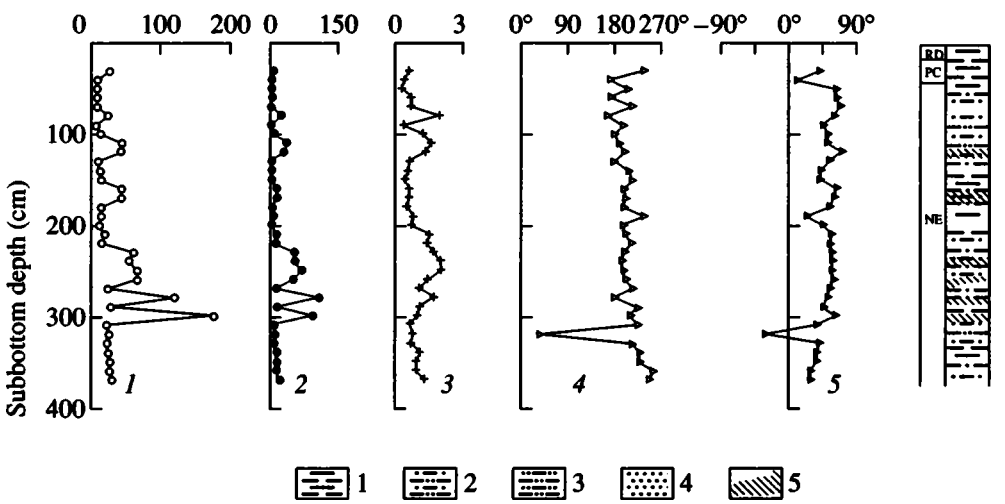

Fig. 5. Paleomagnetic characteristics of deposits recovered at Site 911: (1) pelitic ooze; (2) aleuropelitic ooze; (3) pelitic silty ooze; (4) beds with sandy admixture; (5) beds enriched in hydrotroilite and greigite; *1*, secular variations of magnetic susceptibility ($\chi \times 10^{-6}$ cgs); *2*, secular variations of remanent magnetization ($I_n \times 10^{-6}$ cgs); *3*, factor Q; *4*, declination of I_n vector after thermal demagnetization; *5*, inclination of I_n vector after thermal demagnetization; (RD) Recent deposits; (PC) Paleochernomorian deposits; (NE) Neoeuxinian deposits.

recurrently studied in parallel sections, the submarine sampling procedure does not offer such a possibility, and this complicates the correct interpretation of the results.

As in the study on the Black Sea sector near Bulgaria (Bol'shakov and Nikolaev, 1993), our analysis of magnetic characteristics (magnetic susceptibility, remanent magnetization, thermomagnetic parameters, and others) also shows that they are dependent on the genetic type of sediments, sedimentation environments, and paleogeographic evolution of the sea. The Black Sea evolution was responsible for changes in the abundance and composition of magnetic fractions in

sediments and, thus, modified their magnetic parameters. This primarily concerns the sedimentary strata enriched in hydrotroilite and greigite. These hydrotroilite interlayers, as they are called, have a higher magnetic susceptibility χ, up to 200×10^{-6} cgs, in contrast to other sediments, where this parameter is less than 30×10^{-6}cgs (Figs. 2–5). The I_n and χ values are interrelated, and, consequently, the former is also dependent on the composition and concentration of magnetic minerals in sediments. The correlation coefficient K characterizing the covariance of χ and I_n values is 0.86 for Site 908, 0.90 for Site 911, 0.76 for Site 902, 0.78 for Site 903, and 0.83 for Site 906.

The primary magnetization In0 (a stable part of In after the thermal demagnetization) is also dependent, in general, on the magnetic susceptibility, although this dependence appears to be rather complex when analyzed in detail, as one can see from data obtained for Site 910 (Fig. 2). If the extreme values characteristic of hydrotroilite interlayers are excluded, the magnetic susceptibility of sediments is relatively homogeneous here and varies from 20 to 30×10^{-6} cgs, whereas the primary magnetization changes remarkably. For instance, the average value of I_{n_0} in the depth interval of 50–415 cm is close to 24×10^{-6} cgs, whereas deeper in the column it significantly decreases down to 0.1–0.5×10^{-6} cgs.

V.A. Bol'shakov, who studied the thermomagnetic parameters of our samples, detected that magnetization of the hydrotroilite interlayers with a high magnetic susceptibility is produced, in addition to magnetite,[2] by another mineral having a blocking temperature of 330–360°C and most likely corresponding to greigite. Varieties of pyrrhotite with similar characteristics are of high-temperature origin. Hence, they should belong, if present, to terrigenous components of sediments, which are usually distributed throughout the sedimentary sequence but not concentrated in local horizons such as the hydrotroilite interlayers in the case under consideration. Volkov (1984), who specially investigated sulfide minerals of the Black Sea sediments, failed to find pyrrhotite among them and identified only pyrite, hydrotroilite, and greigite, the latter as an authigenic component of the hydrotroilite interlayers.

Magnetic susceptibility of sediments varies in accordance with changes in sulfide mineralization. The maximum χ values correspond to sedimentary beds, where sulfides are represented by hydrotroilite and greigite, whereas pyrite-bearing layers display the minimum values of this parameter. This is well evident from Table 1, where we present data of Lyu Lei (1993) about concentrations of various Fe-compounds and magnetic susceptibility values of sediments sampled at Site 903. As one can see from the table, parameter χ is positively correlated with the Fe-monosulfide concentration but shows a negative correlation with the content of Fe-disulfide and inert behavior relative to other Fe-compounds. Since such a strong ferrimagnetic mineral as Fe-monosulfide (greigite) coexists with paramagnetic hydrotroilite, the former seems to be solely responsible for the high χ value of respective sediments.

The paleomagnetic stability S of studied sediments is variable. This parameter representing the ratio of the primary magnetization module to the sum of primary and secondary magnetization modules (Khramov and Sholpo, 1967) varies from 0.1 to 0.95. In our case, the primary magnetization was calculated on the basis of results of the thermal demagnetization, and we came to the conclusion that S values are independent of the sediment types. There were two magneto-lithological groups of samples: sediments with magnetic susceptibility higher than 30×10^{-6} cgs (hydrotroilite interlayers with greigite) and other deposits almost lacking this mineral and having parameters of magnetic susceptibility lower than 30×10^{-6} cgs. As one can see from Fig. 6, which illustrates the covariance of paleomagnetic stability S and magnetic susceptibility χ in sediments at the sites 903 and 911, the first parameter varies irregularly relative to the second, and, consequently, it is independent of the sediment lithology. The only point that should be noted is the greater proportion of samples with an S value higher than 0.5 among deposits with an χ parameter lower than 30×10^{-6} cgs. However, these results differ from the data of Tret'yak *et al.* (1989, p. 77), who reported that the percentage of viscous magnetization in sediments is not higher than 10–20%, and, accordingly, their paleomagnetic stability must be no lower than 0.8.

Factor Q_n is another interesting parameter of the studied sediments. Its values vary from 0.1 to 12, and when they are greater than 2, they mark the hydrotroilite interlayers (Figs. 2–5). Usually, high Q_n values indicate that sedimentary rocks experienced chemical magnetization, and this explanation well fits the hydrotroilite interlayers of the Black Sea, since they are enriched in greigite of authigenic origin. The high Q_n factor is characteristic of the Paleochernomorian deposits (Fig. 4), because they were accumulated under the high hydrogen sulfide concentration in the Black Sea water, and a high degree of chemical magnetization in sediments is quite possible in this case.

Paleomagnetic data show that the magnitude of variations of magnetic declination D in the studied sections is high, frequently more than 60°, i.e., greater than the secular variations. Subsequently, it does not reflect changes in the geomagnetic field. At present, we cannot explain this extreme variability of In vector declination. It may be a consequence of slope sedimentation under the influence of contour currents and other factors already mentioned above. The inclination of the I_{n_0} vector is less dependent on all these factors, the gravity processes included. This parameter better reflects the

[2] V.A. Bol'shakov observed magnetite in all the samples studied despite the presence or absence of hydrotroilite and greigite.

Dependence between the magnetic susceptibility and the content of ferrum forms in the bottom sediments of the Black Sea (along a Site 903 column)

Depth, cm	FeS content	FeS_2 content	Content of reactionary Fe	Total Fe content	$\chi \times 10^{-6}$ cgs
	% of dry substance				
0–10	Traces	1.22	2.05	5.64	7.1
10–20	Traces	0.92	2.07	3.29	7.1
53–55	0.20	0.70	1.99	–	26.0
66–68	0.21	0.10	2.05	5.19	58.0
92–95	No	0.05	1.51	4.69	17.5
152–155	1.18	0.05	2.11	–	96.5
155–160	0.04	0.06	1.50	3.90	26.6
236–240	0.54	Traces	1.40	5.36	48.9
337–342	0.42	0.04	1.57	5.19	143.8
370–375	0.05	0.13	1.45	–	30.0
425–430	Traces	1.63	2.68	5.24	10.0
460–465	Traces	0.86	1.76	5.14	5.8
505–510	Traces	1.26	2.23	5.14	4.8
560–565	Traces	1.26	2.23	5.14	7.0

Note: Gray areas mark hydrotroilite interlayers.

magnetic-field changes, although it is lesser in general than the calculated inclination of the present geomagnetic field in the test area, which is equal to 63°. Nearly all studied sections demonstrate the inclination angle decrease from the top downward, which is related to sediment consolidation at the depth responsible for flattening of I_n vectors. Figure 7 shows that the inclination angle decreases by 20° from the top to the base of core sequences at sites 903 and 911 and almost by 40° at Site 910. The only exception is Site 908, where the inclination angle decreases by 10° down to the depth of 445 cm, but, below this level, its average value ($J = 64°$) is greater than in overlying deposits and close to the theoretical magnitude. The decrease in the inclination angle with depth, especially significant at Site 910, indirectly suggests that magnetic particles of sediments are elongated (or aggregated into chains), and, thus, their orientation can be easily affected by the slope sedimentation factors, e.g., by contour currents.

Core sections at sites 903, 908, 910, and 911 comprise intervals of magnetic anomalies, where declination and inclination of I_{n_0} vectors concurrently change. For instance, at Site 903 (Fig. 3), the inclination reversal ($j = -28°$) at the depth of 490 cm is accompanied by the declination rotation through 130°. At Site 908 (Fig. 4), samples from the depth levels of 540 and 545 cm have negative inclination angles of –74° and –67°, respectively, and declinations turn through 130°–140° relative to their directions in the higher and lower horizons. At Site 911 (Fig. 5), the reversed inclination at the depth of 320 cm coincides with the D-vector rotation by almost 180°, whereas at Site 910 (Fig. 2), below the depth level of 480 cm and down to the base of the section, there are irregular fluctuations of declination and inclination of the I_n vector. Moreover, one sample from the depth of 490 cm recorded the inclination reversal ($j = -28°$) at the last site.

Thus, anomalous changes of remanent magnetization vectors and even magnetic reversals are recorded at four above-mentioned sites. It is remarkable that all magnetic anomalies are confined not to the hydrotroilite interlayers but to other sedimentary horizons presumably lacking authigenic minerals and free of chemical magnetization. This allows us to assume the geophysical origin of the observed magnetic anomalies and to consider them as marking the Gothenburg magnetic excursion, because they all are at the level of the Neoeuxinian beds spanning the appropriate age interval.

In other cases (sites 902 and 854), the detected inclination anomalies in the Neoeuxinian beds are not accompanied by substantial changes in declination. These anomalies were not interpreted as a result of the magnetic excursion for the reasons explained above, although this interpretation is also not inconceivable.

It is important to elucidate the stratigraphic position of the recognized excursion relative to the hydrotroilite horizon, as the latter marks one of the most prominent paleogeographic events—the commencement of the permanent inflow of Mediterranean waters into the Black Sea and drastic changes in sedimentation environments (Strizhov *et al.*, 1990). There are two sides to this problem. On the one hand, the presumable isochronism of the excursion offers the datum level for

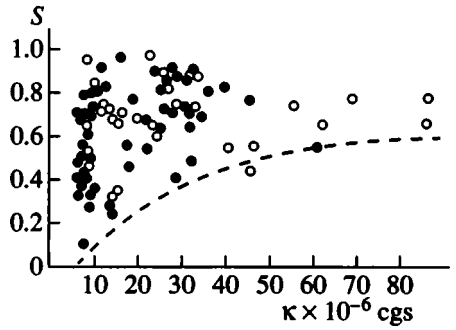

Fig. 6. Covariance diagram of paleomagnetic stability S and magnetic susceptibility χ (open circles for Site 911; solid circles for Site 903).

establishing age intervals of hydrotroilite beds in the western and eastern parts of the Black Sea, where they can be either concurrent or not. On the other hand, we will be able to refine the stratigraphic and chronological position of the excursion, because it is well substantiated that the ingress of Mediterranean waters started at the time of 13 ka.

Unfortunately, Tret'yak *et al.* (1989) did not report about the hydrotroilite horizon in the sections they studied. From this viewpoint, their stratigraphic interpretation of paleomagnetic results seems doubtful.

In the western part of the Black Sea, the magnetic excursion is detected near the base of the hydrotroilite horizon (Bol'shakov and Nikolaev, 1993). Its position in the test area under consideration is less distinct, because the structure of the hydrotroilite horizon is more complex here. However, as one can see from Figs. 2–5, the excursion at sites 903, 908, 910, and 911 is confined to either the base or the lower part of the section interval comprising the hydrotroilite interlayers.

This position of the Gothenburg excursion relative to the hydrotroilite beds suggests their synchronism in the western and eastern areas of the sea. Recently, the paleogeographic evolution of the Black Sea basin and sedimentation factors responsible for the accumulation of hydrotroilite beds were discussed by Strizhov *et al.* (1990), who analyzed the composition of the Neoeuxinian fauna and came to the conclusion that the basin of that time was not of the pure limnetic type and had a salinity equal to 2–3%. Changes in molluscan assemblages show that the water salinity began to increase at the boundary between the Neoeuxinian and Bugazskii horizons, and, by the beginning of the Dzhemete time, it was already close to the current state. These data suggest that the Neoeuxinian–Bugazskii boundary marks the commencement of the Mediterranean transgression.

The depicted secular variations of sulfur, carbon, and oxygen isotope composition (Strizhov *et al.*, 1990) add some details to the evolutionary trend outlined above. The first short-term ingress of Mediterranean waters happened about 15 ka at the end of the late Pleistocene. In the western halistasis, this event is marked by the increased concentration of sulfate, pyrite, and monosulfide sulfur in sediments and by relevant changes in the sulfur isotope composition. The event soon ceased, after a few hundred years. The next transgression, which started about 13 ka ago, lasted until the present time. Initially, the water influx was insignificant; however, it triggered the sulfate-reduction activity of microorganisms and accumulation of hydrotroilite beds. The character of sulfate-reduction processes changed at the end of the Neoeuxinian time, and afterward, environments gradually approached the current state. The hydrosulfuric poisoning now characteristic of the basin commenced in the Kalamitskii interval of time. The sulfate-reduction processes were responsible for the authigenic, almost synsedimentary, formation of

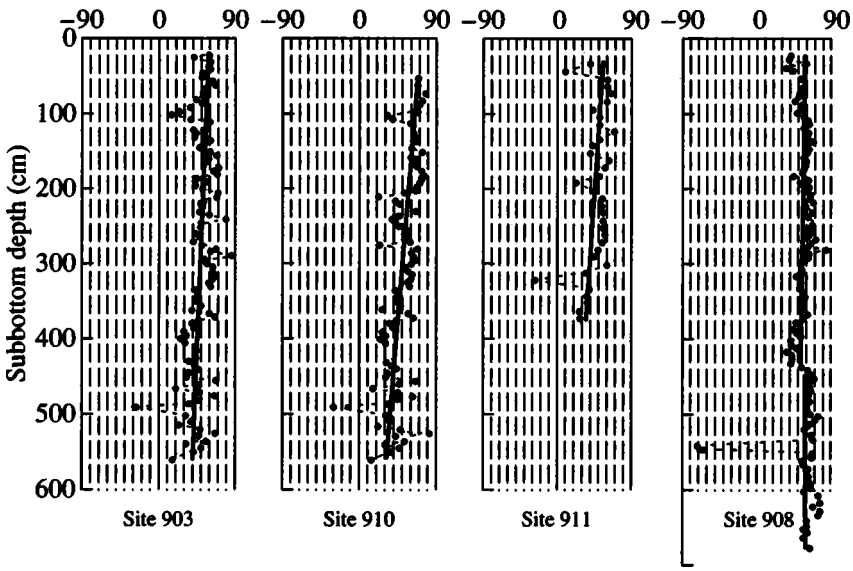

Fig. 7. Variation trends of In vector inclination down the core section at sites 903, 908, 910, and 911.

Fe-sulfide and also for the reduced sedimentation environment.

All these data support our inference based on paleomagnetic measurements and show that the hydrotroilite horizon is almost synchronous all over the Black Sea bottom.

The Neoeuxinian transgression also triggered the readjustment of equilibrium profiles in the shallow-water zones and activated the gravity factors of sedimentation, first of all, the turbidity currents and an intense influx of clastic materials from the newly flooded shelf areas. In the eastern part of the sea surrounded by the Caucasus Mountains, these events were more significant than in the west. This contrast in environments appears to be responsible for the different structure of hydrotroilite beds, which form a single uniform horizon in the west but are separated by other interlayers in the east, where the latter correspond to episodes of an intense influx of sedimentary material from the upper oxidation zone. Material redeposition, presumably recurrent, was also typical of the last region. Despite all the differences, the eastern hydrotroilite horizon, whether it is homogeneous or consists of a series of separate thin interlayers rich in hydrotroilite, must be coeval or insignificantly younger than the western one. This is proved by its position relative to the Gothenburg excursion described above.

CONCLUSION

This study and published data show that the paleomagnetic anomaly in the upper Pleistocene deposits of the Black Sea appears to have been caused by geophysical factors. Its position in the Neoeuxinian beds allows us to identify this anomaly as the geomagnetic Gothenburg excursion. In both the western and eastern parts of the sea, the excursion is confined to the base of the reference hydrotroilite horizon, and this implies the nearly concurrent deposition of hydrotroilite interlayers in all the Black Sea areas.

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Quaternary Stratigraphy and Evolution of the Middle Volga and Lower Kama River Basins

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Abstract—Recent deposits representing the early–late Pleistocene climatoliths are detected and described in the course of a comprehensive study of several reference sections in the middle Volga River basin. Data of lithological, geochemical, paleopedological, and paleontological analyses show that the Quaternary history of the middle Volga and lower Kama river basins is divisible into thirteen paleogeographic stages corresponding to seven interglacial or interstadial events (early to late Il'inka beds = Cromerian II and III; Belovezh'e Horizon = Cromerian IV; Likhvinian Horizon s. str. = Inzhavino beds = Holsteinian; Kamenka, Romny, and Mikulinian beds = Eemian; Bryansk Horizon = Holocene) and to six glacial phases separating them (Don, Okaian, Borisoglebsk, Orchik, Dnieperian, and Valdayan glaciations). The recognized climatic rhythms of the Pleistocene are correlated and used as a basis to elucidate the evolution of soil horizons and natural environments in the East European region of Russia.

Key words: *Recent deposits, loess–soil formation, paleosoils, stratigraphy, paleogeography, correlation, microteriofauna.*

A comprehensive study of recent deposits is still the main concern of Quaternary geology, because the detailed stratigraphic schemes of the Pleistocene and respective paleogeographic reconstructions provide the necessary scientific background for geological survey and prospecting. During recent decades, investigations of this kind were performed in the glaciation and periglacial areas of the middle Volga River basin, where the following Pleistocene reference sections were studied: the Tigany, Zaplatino, Pavlovo, Varezh, Barmino, Gorodets, Rybnaya Sloboda, Tatarskaya Chishma, Rechnoe-Ostolopovo, and Komintern sections, all located in the lower courses of the Oka and Kama rivers, on the left-hand side of the Volga River; the Chirkovo, Kuchenyaevo, and Kivat sections distributed along the Sura River; the Attikovo and Apastovo sections at the Sviyaga River; the Berezovka section at the Tesha River; the Dubenskoe section at the P'yana River; the Shakhmanovo and Barannikovo sections at the Sundovik River; the Elasy section at the Uynga River; and the Krasnye Baki section at the Vetluga River (Fig. 1). The newly obtained data provided insight into the geological structure of the Pleistocene deposits and were used in paleogeographic reconstructions and comparative analysis of main environmental factors. They also elucidate some stratigraphic problems and cast light on the accumulation environments of recent deposits and on the paleogeographic history of the investigated areas.

Data available at present show that the stratigraphy of subaerial deposits and paleontological characteristics of subaquatic sedimentary beds are poorly understood and rather disputable. At the same time, the loess–soil formation widespread in the region under discussion is very informative with respect to paleogeographic events that happened over a period longer than 1 m.y., and particularly with respect to alternating interglacial and glacial epochs responsible for respective periods of pedogenesis and loess accumulation, both representing the most characteristic features of the Pleistocene geology (Markov and Velichko, 1967; Velichko, 1973). The loess–soil formation bearing remains of microteriofauna is the key object comprising those recent deposits, which are important in stratigraphic and correlation aspects as indicators of paleogeographic events. Their significance in the Quaternary climatostratigraphy was well demonstrated by Shantser (1935) and Moskvitin (1958, 1965), who successfully used paleosoils in the stratigraphic subdivision of deposits in the middle Volga River basin.

Stratigraphy and chronology of subaerial deposits discussed below is also based on historical stages of development of the loess–soil formation, which is unique in typology, reveals a series of diagnostic features, and comprises soil horizons diverse in origin and representing the geochronological reference levels of the Pleistocene. These levels are recognized in various sections and successfully traced over the studied region, since the soil formations were genetically interrelated in the past. The loess–soil sedimentary beds constitute up to 80% of the Pleistocene sequence. The detected genetic interrelations between these beds inside the middle–late Pleistocene interval and variable

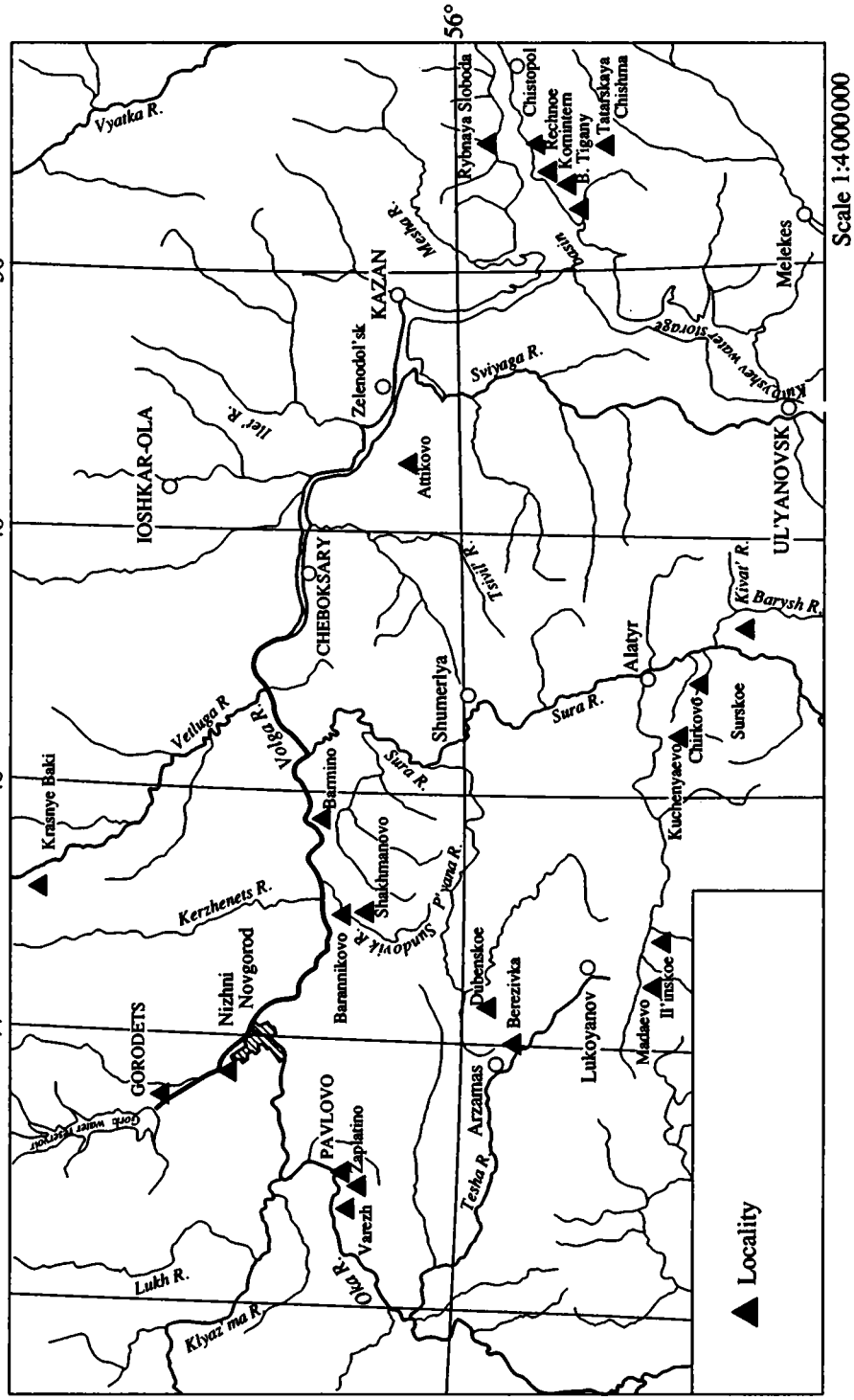


Fig. 1. Localities of studied Pleistocene–Holocene sections in the middle Volga and lower Kama river basins.

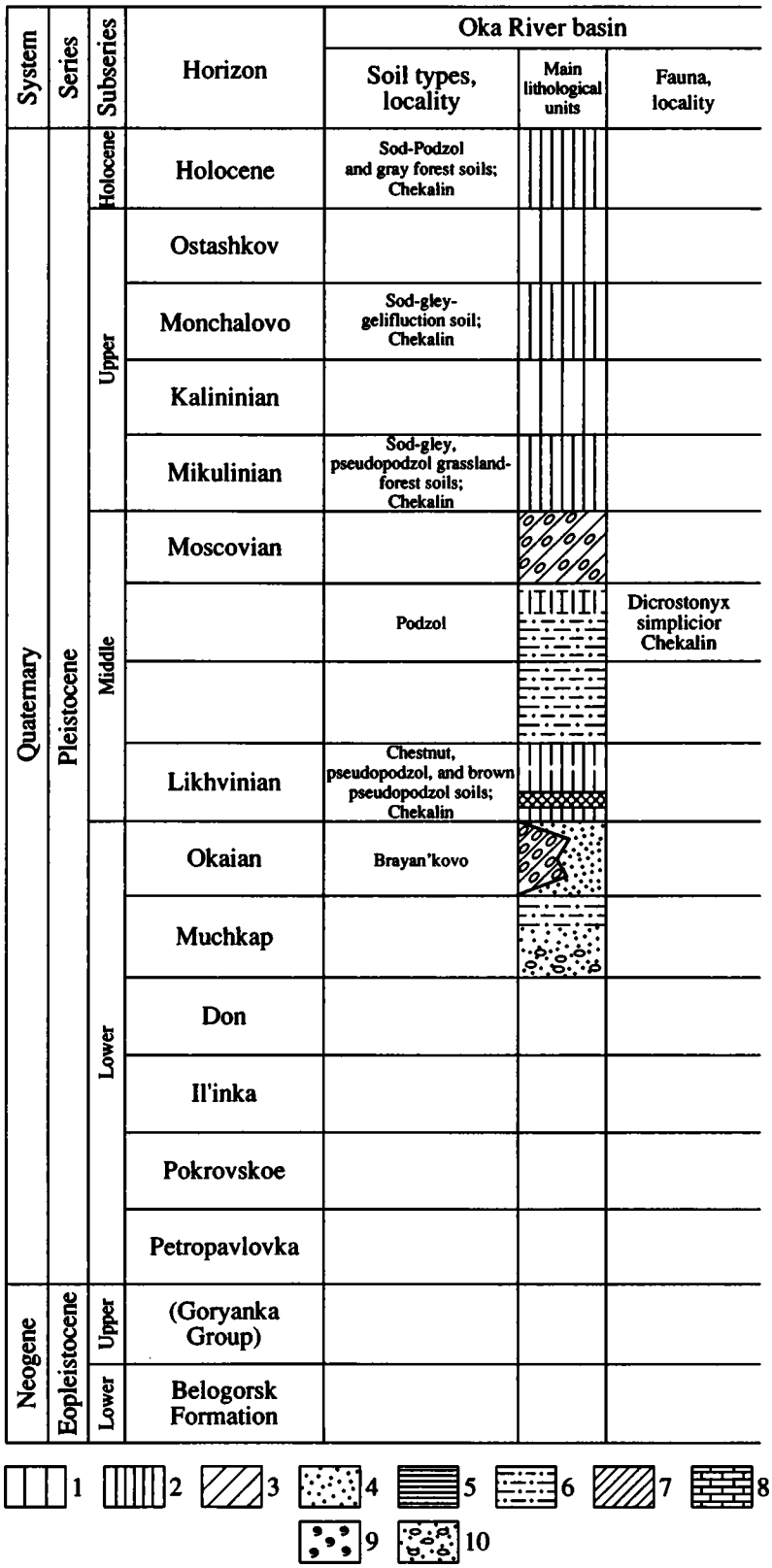


Fig. 2. Stratigraphy and correlation of recent deposits in central and eastern areas of the Russian plain: (1) present-day and fossil soils; (2) loess-like loam; (3) moraine; (4) sand; (5) clay; (6) silt; (7) gyttja; (8) limestone; (9) molluscan shells; (10) pebble bed.

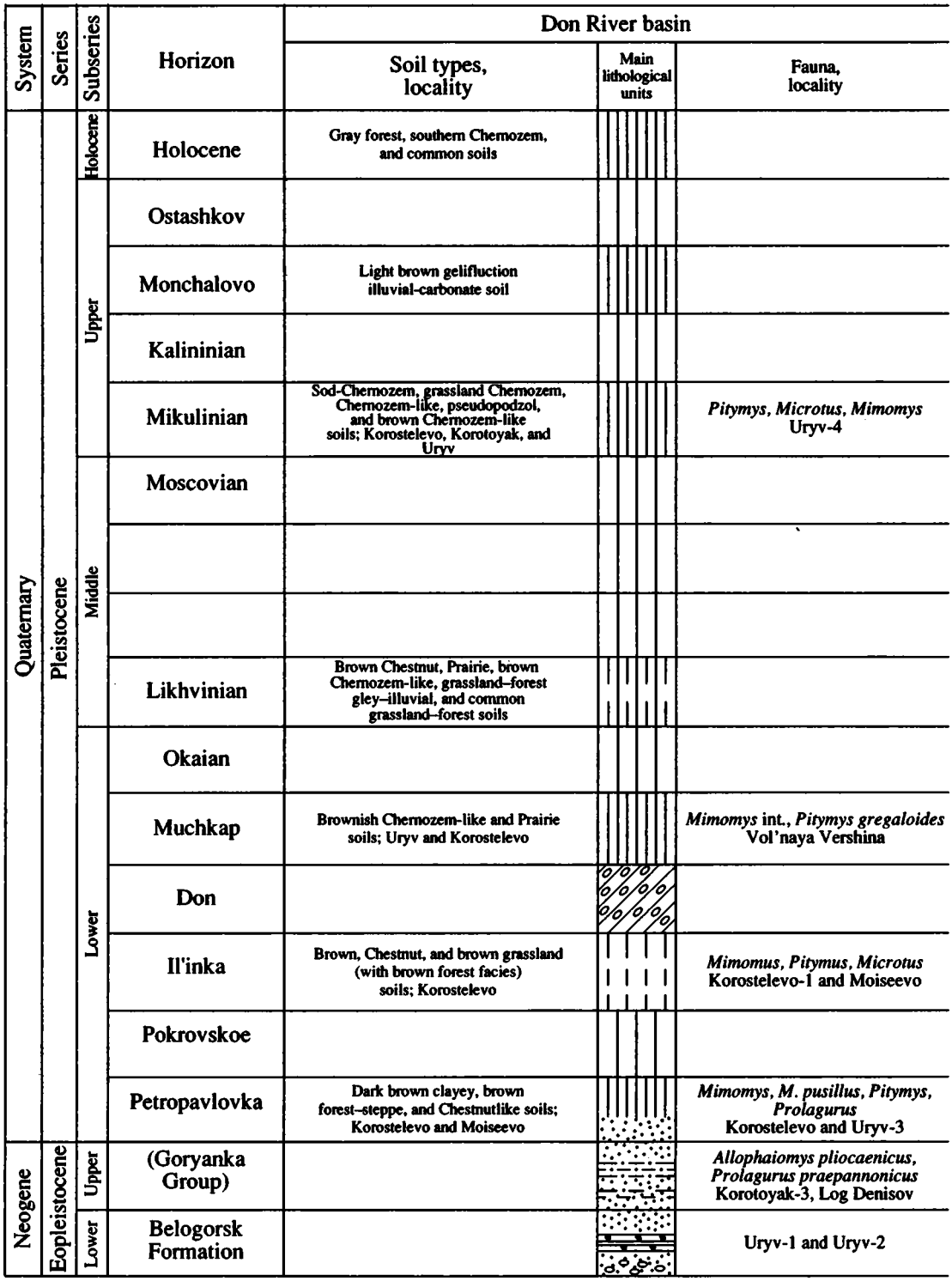
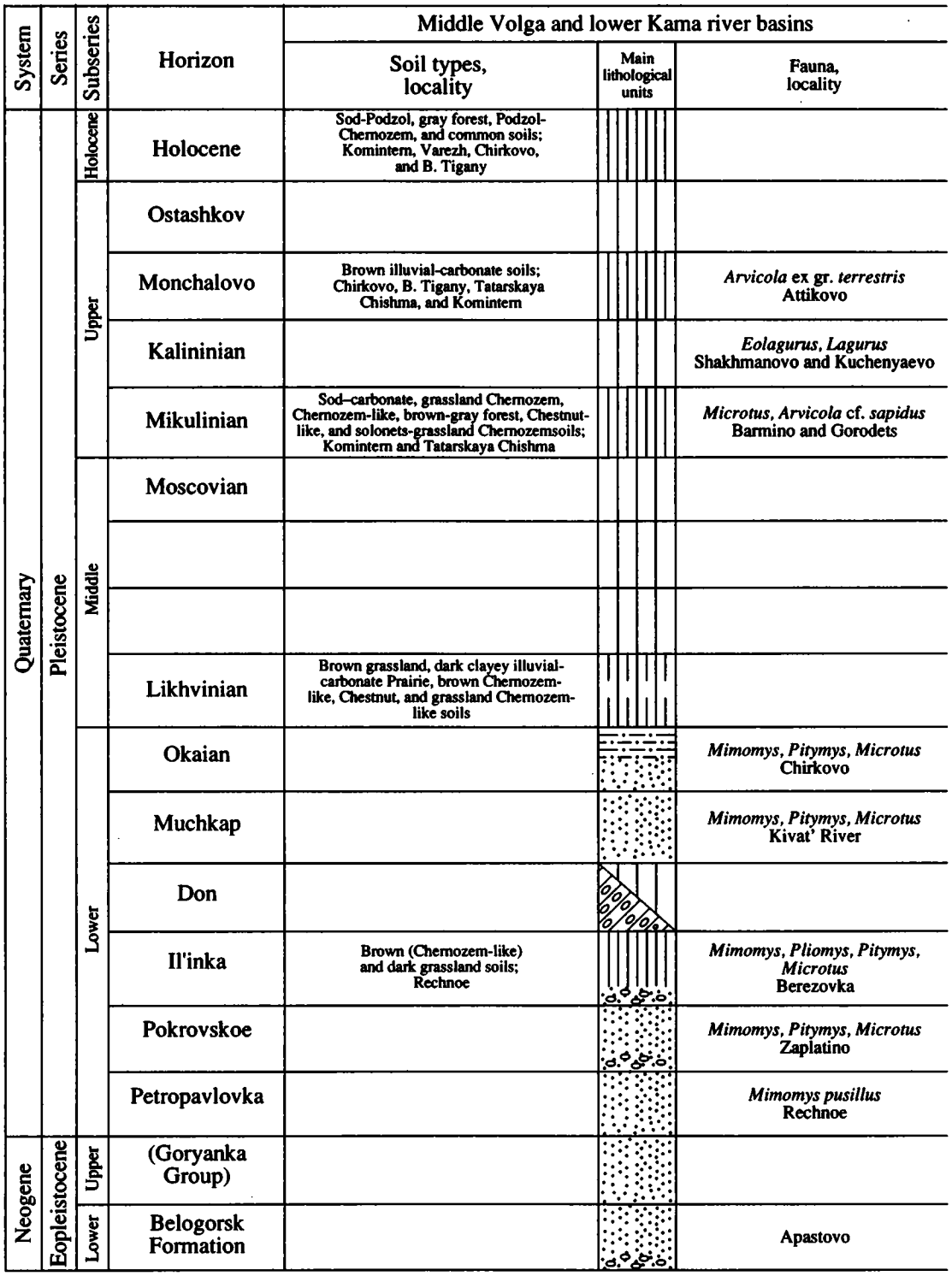


Fig. 2. (Contd.)

facial transitions from them and to alluvial and glacial deposits provided the necessary basis for the local and regional correlation. These characteristics also revealed a peculiar structure of subaerial deposits in the middle Volga River basin and specific time- and space-dependent trends of soil development and loess accumulation.

Terrestrial Pliocene deposits underlying the Pleistocene sequence consist of sand and less abundant clay. A special investigation in several localities (Apastovo,



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Karan-Azikovo, Daukovo, Yunga-Kushurga, and others) added many details to the former understanding of their structure, lithology, and paleontology. The Apastovo reference section is one of the best studied. It is located at the watershed plateau, where alluvial and lacustrine–alluvial deposits (6.2–9.9 m thick) of the ancient Sviyaga River are exposed. The upper part of the sequence (4.3–6.5 m thick) is composed of sand beds with plant remains, molluscan shells, and fossil bones of mammals. It rests on the erosion surface of

gray to dark gray clay beds also bearing rare molluscan shells and mammalian bones (Glushankova, 1992).

The bone collection of small mammals from the upper alluvial formation of the Apastovo locality consists of several thousand specimens. A.G. Agadzhanyan (Paleontological Institute, Russian Academy of Sciences), who studied this fauna, has concluded that lagomorphs of the *Hypolagus* genus represent about 20% of its composition. The group of less abundant insectivores includes *Blarinoides* and *Desmana* forms associated with the small mole species *Talpa minuta* Blainv. Voles *Mimomys polonicus, M. hintoni* Fejfar, and *Promimomys baschkirica* Suchov are most frequent among rodents. In contrast, *Villanja* (*Borsodia*) *veterioe* Kretzoi, small beaver, rhyzodont zokor *Prosiphneus*, and the lemming *Lemmini* represent the rarely occurring Rodentia forms. In the taxonomic composition and evolutionary level of rodents, the Apastovo fauna resembles the Simbugino and Akkulaevo faunas from the Akchagylian deposits of Bashkortostan (Sukhov, 1970, 1977). Consequently, the upper Pleistocene beds of the Apastovo locality are correlative with the Akchagylian strata of the Cis-Urals region. The fauna in question is also comparable with assemblages of small mammals from the Korotoyak-2 and Uryv-1 localities at the Don River (Fig. 2) and from the Hainachka, Archille, Gundershaim-4, Baleru-2, and Charnota-2 localities of Western Europe. In the ecological features, this fauna corresponds to a community of forest and forest–steppe biotopes that existed under conditions of a warm and sufficiently humid climate. In addition, deposits with similar microteriofauna are known at the Elasov (the Yunga River basin) and Naberezhnye Chelny (the left-hand side of the Kama River) localities. This suggests that paleogeographic inferences with respect to the Apastovo locality can be valid for the entire region of middle courses of the Volga River (Agadzhanyan and Glushankova, 1989).

Early Pleistocene. The lower Pleistocene deposits in the studied region are represented by fragments of a well-developed soil horizon of the interglacial type and also by alluvial beds in the buried valleys of the Oka, Volga, Kama, Sura, and Tesha rivers. The alluvial beds are 1.6 to 4.6 m thick and quite uniform in lithology. They display either horizontal or crossed bedding and consist of alternating yellow-gray sand, gray to dove-colored clay, and brown to gray-brown loam. In many places, the beds yielded fossil bones of small mammals of the Tiraspol assemblage. The oldest alluvial deposits were detected in the Zaplatino section on the right-hand side of the Oka River (Fig. 1). Bone remains of microteriofauna were discovered here in the lower bed (2.55 m thick) of light gray to yellow sand enclosing gravel, pebbles, and abundant shells of freshwater mollusks. This fauna mainly consists of voles: rhyzodont *Mimomys* and arhyzodont *Pitymys hintoni gregaloides* associated with *Microtus* forms resembling the present-day voles from northern Siberia. Sagebrush voles are also rather abundant. Lemmings, bank voles, ground squirrels, haymakers, insectivores, and lagomorphs are the accessory components. The fauna and host rocks are the early Pleistocene in age. Most archaic arhyzodont voles are missing here, and their rhyzodont forms are not diverse. The last features imply that the rodent community under consideration is younger than the late Taman fauna of the Apcheronian time known from many sites: the Demsk and Davlekaevo localities of Bashkortostan; the Uspenka, Denisov Log, and Korotoyak-3 localities at the Don River; and the Nogaisk, Tarkhankut, Tiligul, and Chishmiknoi localities in the Black Sea region. At the same time, the Zaplatino fauna is more archaic and older than typical late Tiraspol faunas from the following localities: Chui-Atasevo in the Cis-Urals region; Novokhopersk, Bogdanovka, Uryv-4, Moiseevo-3, and Kuznetsovka at the Don River; Roslavl at the Desna River; Korchevo in Belarus; and Kolkotova Balka at the Dniester River. The Il'inka and Veret'e faunas from the Don River basin and the Platovo fauna from the Sea of Azov region were found to be the most precise age analogues of the Zaplatino assemblage (Agadzhanyan and Glushankova, 1987, 1988). The latter is also close to faunas from the Feugstadt beds of Germany and Cromerian beds of England.

The fauna is peculiar in the ecological aspect. A high abundance of bush (*Pitymys*) and bank voles (*Clethrionomys* forms) indicates that coniferous–broad-leaved forests were rather widespread at the existence time of the fauna. The respective landscape also included local steppe areas free of arboreal vegetation favorable for rather abundant lagurids and coexisting ground squirrels and haymakers. These areas should have some components of the fruticose tundra biota, as follows from the presence of collared lemming. The gray lemming forms suggest another component of the landscape, such as mosseries in the relief depressions. The respective climate was notably cold and dry; however, the adjacent land was free of glaciation. Thus, the Zaplatino fauna marks one of the cold to moderately cold epochs of the early Pleistocene. Paleogeographic environments of that time were responsible for peculiar lithological and geochemical features of alluvial sediments exposed in the Varezh section and identical to deposits of the Zaplatino locality. Wide ranges of indicative ratios (SiO_2/Al_2O_3 = 6.63–13.09, SiO_2/Al_2O_3 = 20.41–43.95) show that sediments of the former locality did not experience a considerable retrodiagenesis. The alluvial beds here have low concentrations of the fine-grained fraction (5.9–8.5%), carbonate material ($CO_{2\,carb}$ = 0.12%), organic matter (C_{org} = 0.10%), and easily soluble compounds (dry residue 0.035%).

In the Volga Upland, like everywhere in the Russian platform, the incision of the river system and alluvium accumulation in the ancient river valleys were initiated at the beginning of the early Pleistocene (Goretskii, 1964). The lower Pleistocene alluvial deposits (3.2 m thick) of the Volga Upland were studied in the Sura

River basin (Agadzhanyan *et al.*, 1988). In the Chirkovo section located here (Fig. 1), the basal bed rests on Cretaceous sand and is composed of pebbles and gravel cemented by the yellow-brown matrix consisting of sand and loam. The bed yielded bone remains of small mammals, such as *Mimomys pussilus* Mehely, *Pitymys gregaloides,* and *Prolagurus pannonicus,* whose assemblage is characteristic of the initial half-period of the early Pleistocene. This fauna is correlative in age with faunas from the Zaplatino locality at the Oka River and from the Il'inka and Veret'e localities in the Don River basin, as they all incorporate the rhyzodont (*Mimomys*) and arhyzodont (*Pitymys, Microtus*) vole forms. The coexistence of *Pitymys, Prolagurus,* and Siberian vole forms is an indicator of moderate climatic conditions. Landscapes of that time were presumably rich in both the forest and the steppe biotopes and also in some boreal components.

Alluvial deposits of the Berezovka locality on the right-hand side of the Tesha River (Fig. 1) correspond to a warm stage of the initial early Pleistocene. These deposits underlying the morainal bed (3.5–5.2 m thick) are composed of pebble–gravel material embedded into the gray-brown sand or loam matrix; they also include cross-bedded interlayers of medium- and coarse-grained sand (1.95–2.85 m thick). The bone-bearing horizon of the locality is rich in shells of freshwater gastropods. The archaic rhyzodont *Mimomys* and *Pliomys,* along with more progressive arhyzodont *Pitymys* and *Microtus,* are the main taxa of the Berezovka fauna. Other rodents are represented by ground squirrels, mole rats, hamsters, bank voles (*Clethrionomys*), and field mice (*Apodemus silvaticus*). Insectivores, both small (screws of *Sorex, Drepanosorex,* and *Beremendia* genera) and large (muskrats and moles), are more abundant and diverse here than in other localities. Lagomorphs, mainly piping hares, are not abundant. As noted before, the preponderant *Mimomys, Pitymys,* and *Microtus* forms define the early Pleistocene age of the fauna in question, and this inference is affirmed by the presence of *Pliomys episcopalis* and also by the morphology of muskrat forms. The latter display features characteristic of the early Pleistocene species *Desmana magna* Owen and *Desmana ternopolitana* Pidoplicko. In general, the Berezovka fauna looks somewhat younger than the Zaplatino fauna described above and younger than the Novo-Troitskoe, Veret'e, and Il'inka faunas from the Don River basin. At the same time, it is older than the terminal early Pleistocene faunas of the Kuznetsovka, Vol'naya Vershchina, and Korotoyak-4 localities in the last region. The closest analogues of this fauna are known from beds underlying morainal deposits at the Novokhopersk, Moiseevo-2, Moiseevo-3, Korostelevo-1, and some other localities of the Don River region (Agadzhanyan and Glushankova, 1987, 1988).

Species predominant in the fauna composition, such as field mice, flying squirrels, and bank and bush voles, belong to the forest community. Steppe forms are not abundant and less diverse, whereas boreal and arctic taxa are missing. Such a composition of rodents is characteristic of biotopes from coniferous–broad-leaved forests. The abundance and diversity of insectivores also point to favorable paleogeographic environments of this type. Desmans of the assemblage suggest that water basins of the past landscape were rich in mollusks and other invertebrates. These basins probably did not freeze down to the bottom even in their shallow zones. Simple moles and mole rats again suggest a mild climate excluding freezing of the ground down to a considerable depth in the winter time. The studied region of the Volga River basin was presumably rich in coniferous–broad-leaved forests and grassland biotopes in the floodplain areas populated by sedge and gramineous taxa. Areas with steppe assemblages were not abundant. The landscape had a well-developed river system and many, though shallow, water basins rich in biota. Thus, the climate was warm and moderately humid, corresponding to an interglacial optimum, and alluvial beds below the morainal deposits of the Berezovka section should correspond to the Il'inka Horizon of the regional stratigraphic scheme of 1984 approved by the Interdepartmental Stratigraphic Committee (ISC). Data obtained earlier in the Don River basin show that recurrent climatic changes during the Il'inka time resulted in accumulation of three alluvial complexes bearing the early Tiraspol microteriofauna and thermophilic mollusks. These complexes are separated by fossil soil horizons (Agadzhanyan and Glushankova, 1989; Glushankova, 1990a, 1990b, 1990c, 1992, 1994).

The well-developed paleosoil of the early Pleistocene (2.0 m thick) was formed during the late Il'inka interstade and corresponds to oxygen-isotope excursion 17 (Shackleton and Opdyke, 1976). This soil horizon was recovered by prospect trenches at the surface of the 4th terrace above the flood plain (the Rechnoe-Ostolopovo locality of the Kama River bank near the town of Chistopol in Fig. 1). Its age is determined on the basis of the early Tiraspol microteriofauna from the underlying and overlying beds. The soil profile is differentiated into the brown humic (0.6 m thick) and light brown illuvial horizons. A detailed study showed that sod-forming processes in the Kama River basin were very active during the late Il'inka interstade (C_{org} = 0.11–0.47%; C_{ha}/C_{fa} = 1.0–1.2). Additional soil-formation factors included the chemical migration of main soil components (SiO_2/Al_2O_3 = 2.97–3.40; SiO_2/Al_2O_3 = 4.07–4.73; SiO_2/Al_2O_3 = 11.05–12.65), including carbonates ($CO_{2\,carb}$ = 1.32–5.13%), along with gleying and calcium carbonate accumulation in the lower part of the soil profile. The clay content increases in the lower–middle portion of the humic horizon and in the upper part of the illuvial horizon (31.7–41.0% of fraction <0.001 mm), which means that soil formation took place under warm conditions in a well-developed hydrogenic zone of the past landscape. Morphological and geochemical features suggest this soil to be of the grassland Chestnut (Cherozem-like) type. Soil com-

plexes varying from the dark grassland (Chestnut) to Cherozem-like types presumably were formed in the relief depressions of the terrace, where the reduction–oxidation state was governed by a close horizon of weakly mineralized groundwater of calcium hydrocarbonate composition (Glushankova, 1990b, 1992).

Alluvial deposits at the base of the Rechnoe-Ostolopovo section (1.5 m thick), which are parental for the soil, consist of brown silt and light brown fine-grained sand. They yielded remains of *Sorex* sp. (1), *Mimomys pussilus* Mehely (12), *Microtus oeconomus* Pallas (4), *Microtus* sp. (12), *Citellus* sp. (1), and *Lagurini* gen. (1), suggesting that sedimentation took place in a small temporary basin during a warm and relatively arid climatic phase. This inference is supported by the concurrent fauna of mollusks, among which M.F. Ivakhnenko (Paleontological Institute, Russian Academy of Sciences) detected *Planorbis planorbis* (L.), *Limnaea* (*Stagnicola*) *vulnerata* Küster, *Pupilo miscorom* (L.), *Anisus spirobis* (L.), and *Anisus leuciistoma* (Millet.). The brown loam–sand beds overlying the fossil soil were deposited in the flood plain, where conditions were similar to those described above. The bed yielded microteriofauna of *Mimomys* ex. gr. *pussilus* Mehely and *Prolagurus pannonicus* Kormos associated with mollusks *Valvata ambigna* Westerlund, *Anisus dazuri* (Morch), *Succinella oblonga* (Draparhaud), and *Musculium creplini* (Dunker). The first and second of the listed species indicate the early Pleistocene age of their host sediments.

Alluvial deposits of the terminal half-period of the early Pleistocene were studied at the base of the 4th terrace above the floodplain known as the Kivat' locality of the Barysh River. Most abundant taxa of the Kivat' fauna are represented by rhyzodont *Mimomys* and arhyzodont *Pitymys* and *Microtus* forms. They coexisted with vole *Clethrionomys,* mole rat *Spalax,* beaver, ground squirrel *Citellus*, screw, and rather abundant lagomorphs of the *Ochotona* and *Lepus* genera. Several features, such as the coexistence of rhyzodont and arhyzodont voles, the evolutionary level of *Pitymys* forms, the presence of narrow-cranium voles, and the abundance ratios of *Pitymys gregaloides* Hinton to *Pitymys arvaloides* Hinton and *Pitymys* to *Microtus* species, show that this rodent community originated in the second half of the early Pleistocene and corresponds to the evolved late Tiraspol faunas. The assemblage composition implies variable paleogeographic biotopes. *Clethrionomys* and *Pitymys* forms are characteristic of forest areas, whereas *Spalax, Citellus,* and *Stenocranius gregalis* species are typical representatives of steppe and forest–steppe biotopes. They all coexisted with beaver and root mole, who were dwellers of past river valleys with underwood, grassland, and sedge-bog areas. The data obtained lead to the conclusion that the described alluvial deposits of the Barysh River were accumulated during a warm epoch of the second half-period of the early Pleistocene. The Kivat' fauna is younger than the Berezovka, Zaplatino, and Chirkovo faunas. It correlates best of all with the Moiseevo-2 and Moiseevo-3 faunas from the Don River basin (Agadzhanyan and Glushankova, 1989). According to its interglacial aspect, the fauna may be concurrent with the Muchkap Horizon of the regional scale of 1984 accepted by ISC. In addition to the Don River basin, deposits of the Muchkap (Roslavl) Horizon, which bear late Tiraspol microteriofauna in the stratotype and other localities and correspond to two climatic optimums separated by a significant cooling epoch, are known in Belarus, Poland, Germany, and Holland (Mojski, 1985; Zagwijn, 1985; Agadzhanyan and Glushankova, 1989; Shik, 1995). The horizon is coeval with the soil complex that is complicated in structure and occurs in the Korostelevo section among subaerial deposits of the Don River basin, and also with the Paleoeuxinian transgression in the Black Sea basin. The interglaciation in question likely corresponds to oxygen-isotope excursions 15, 14, and 13.

Deposits of the terminal early Pleistocene in the middle courses of the Volga River are represented by the Chirkovo sequence of brown and bluish dove-colored loam beds (1.8 m thick in total) of lacustrine-periglacial origin, which were accumulated outside the glaciation limits. Lithological and geochemical characteristics of these beds imply peculiar environments of sedimentation, as the rocks are almost free of carbonates ($CO_{2\,carb}$ = 0.10–0.15%), humic matter (C_{org} = 0.15–0.21 %), and easily soluble salts (dry residue 0.040%) and are weakly altered by retrodiagenesis into the loess-like substance. The beds overlie the alluvial deposits bearing the Tiraspol microteriofauna and may correspond to the Okaian glaciation (Agadzhanyan *et al.*, 1988; Glushankova, 1992). Their analogues, most likely correlative with oxygen-isotope excursion 12, are known in many sites, e.g., in the Don and Dniester River basins.

Middle Pleistocene. Deposits of this age are exposed in the Chirkovo, Tatarskaya Chishma, Komintern, and other sections of the Volga middle courses (Fig. 1). Available geological, paleopedological, and paleontological data, along with results of direct chronostratigraphic correlation of concurrent reference levels, show that the loess–soil formation of the Okaian–Dnieperian time is divisible into three interglacial horizons of fossil soils separated by loess beds. The soil horizons mark three epochs of intense pedogenesis during the middle Pleistocene successively corresponding to the Likhvinian and Kamenka interglacials and the Romny interstade (Glushankova, 1992, 1994).

A complete profile of the paleosoil (pedocomplex), though considerably deformed, is known at the base of the middle Pleistocene sequence. It was formed during the Likhvinian interglacial, because the underlying alluvial deposits yielded the well-known Singil' microteriofauna (Velichko *et al.*, 1989; Markova, 1992). At the optimum of the Likhvinian interglaciation, the ground of forest and forest–steppe landscapes in the

middle Volga and lower Kama river basins mainly consisted of brown soil complexes similar to the Central European soil facies. The same Cherozem-like soil (in terms of Antipov-Karataev, 1932) or the East European Brunizem (in terms of I.V. Tyurin) are known in the present forest–steppe zone adjacent to the Kama River. They represent a certain stage in evolution of the recent soil horizon that marks a transition from the Chestnut soil to the Cherozem according to all diagnostic criteria (Liverovskii, 1948). The pedocomplex profile includes the crimson-brown accumulative humic horizon (0.55 m thick) with mole tunnels and isolated Fe-Mn-microhardpans. This unit grades downward into the light brown ferruginate illuvial horizon (0.7 m thick) with carbonate concretions, dispersed silica grains, and abundant mole tunnels.

Paleosoils resembling the present-day Chestnut (Cherozem-like) soils display several characteristic features: they are highly humic (C_{org} = 0.32–0.82%), deeply colored with organic matter (C_{ha}/C_{fa} = 1.33–3.0), and rather uniform throughout the profile with respect to granulometric parameters, although with depth they become somewhat less heavy. Their chemical composition is nearly constant (SiO_2/R_2O_3 = 5.04–5.60; SiO_2/Al_2O_3 = 6.56–7.39; SiO_2/Fe_2O_3 = 2.18–2.31), but the peak content of carbonates is confined to the illuvial horizon ($CO_{2\,carb}$ = 7.85%). The purification horizon A_2 is missing in the structural succession of the profile, and the fine-grained fraction is insignificantly concentrated below the humic horizon (Glushankova, 1990b, 1992). These characteristics may imply that the soil-formation mechanism of that time included the following elementary processes: humification, humus accumulation, and claying of the profile without a notable migration of weathering and soil-forming products. Morphological and chemical diagnostic criteria point to a complex history of soil development under the grassland–forest and grassland–steppe vegetation of the watershed areas, where sod formation was one of the main factors.

The loess-like clayey loam above the pedocomplex was presumably formed when the soil-formation processes became less active after the climatic cooling and commencement of a more arid phase.

Higher in the middle Pleistocene loess sequence exposed in the Kama and Sura river basins, there is a rather thick (2.5–2.85 m) bed of heavy clayey soil (33.0–42.08% of fraction <0.001 mm). Its polygenetic profile includes a dark gray to light brown humic horizon (1.2 m thick) with abundant Fe-Mn-hardpans. The horizon grades downward into a bluish or grayish brown illuvial horizon with rare spots of Fe-Mn-hydroxides. This paleosoil is rather uniform in granulometric and geochemical aspects. Displaying a considerable degree humification of (C_{org} = 0.26–0.49%), notable carbonate content ($CO_{2\,carb}$ = 3.6–6.4%), and humate composition of humus (C_{ha}/C_{fa} = 0.95–1.5), the soil has a high absorption capacity, especially in the humic horizon (24.6–28.98 mg/ecv). The fine-grained soil fraction is dominant everywhere in the profile. Calcium and magnesium, especially the former, are the principal absorbed cations of the soil. Concentration of soluble compounds is low (dry residue from 0.053 to 0.085%); these are calcium hydrocarbonates in the Komintern paleosoil and sodium sulfates in the soil of the Tatarskaya Chishma locality. In this case, the soil-formation process mainly consisted in accumulation of humus and migration of soluble compounds (sometimes together with the clay fraction) either into the lower part of the profile or outside the soil horizons. The hydromorphic properties of the soil and rich herbaceous vegetation were favorable for the soil saturation with humus and basic ions. Intense atmospheric precipitation presumably was responsible for the periodic leaching of the soil, as is typical of the present-day Chestnut soils, and also for a high level of groundwaters in the watershed areas. The peculiar morphology and other characteristics suggest that this soil is of the Cherozem-like type with relic features of loessification and originated in the grassland areas. Similar soils are known in the Seim and Don river basins (the Mikhailovka, Korostelevo, Uryv, and other localities of paleosoil beds) and also in the prairie areas of the United States and prairielike zones of the Far East, which are now typical of subtropical regions with a moderately continental and moderately humid climate (Glazovskaya, 1973; Liverovskii, 1987; Agadzhanyan and Glushankova, 1986, 1987; Glushankova, 1990a, 1990b, 1990c, 1994). Terminal formation stages of the paleosoil are marked by gelifluction deformations resulted from climatic cooling in a more continental environment. The soil-formation stage at the interglacial optimum presumably corresponded to a high bioproductivity phase, when humidity conditions were more favorable than in typical steppe zones at present.

The next epoch of the middle Pleistocene pedogenesis took place in different paleogeographic environments. In many reference sections (the Chirkovo, Tatarskaya Chishma, Komintern, Rybnaya Sloboda, and others), the Romny paleosoils are divisible into horizons A_1, B_t, B_{Ca}, and C. The soils are confined to the base of the Dnieperian Horizon composed of loess-like clayey loam. The paleosoil level in the Kama and Sura river basins is separated from lower pedocomplexes of the Kamenka interstade by another bed of loess-like clayey loam (0.6–0.7 m thick), which was affected by pedogenesis (Glushankova, 1992, 1994). Microteriofauna recovered from mole tunnels in the paleosoil was referred to the Khazarian assemblages (Markova, 1982, 1992).

In the terminal middle Pleistocene, pedogenesis in the watershed areas resulted in accumulation of a soil horizon with considerably weathered mineral mass, and this feature distinguishes the horizon from both the younger Holocene and older soils. Conditions of pedogenesis at that time presumably were notably warmer. A high biogenic activity, claying processes, and accu-

mulation of carbonates parallel to humification reveal a similarity between these and present-day Chestnut soils occurring in subtropical and warm boreal regions, where the seasonal semiarid climate resembles that of the Mediterranean (Antipov-Karataev and Prasolov, 1932; Liverovskii, 1948; Gerasimov, 1954, 1976; Rozanov, 1957; Glazovskaya, 1972).

Brown paleosoils of the Romny period differ from others in their poorly differentiated clay–carbonate profiles. These profiles are rather uniform with respect to distribution of principal chemical components, and their differentiation is evident only from the claying maximum confined to the middle part. Respective soils in the lower Volga and middle Kama river basins likely represent the steppe analogues of the grassland Chestnut soils. The former are 0.8–1.3 m thick, gray-brown in color, and highly humic (C_{org} = 0.25–0.57%). They belong to the calcium-humate type (C_{ha}/C_{fa} = 1.1–1.3) and display a rather high carbonate content ($CO_{2\,carb}$ = 4.28–5.04%) and better cation-exchange capacity, as compared to coeval soils in the Seim and Don river basins (Agadzhanyan and Glushankova, 1986, 1987; Glushankova, 1990a, 1990c, 1992).

Morphology and physicochemical characteristics suggest that paleosoils of the last type were formed above deep-seated groundwaters without a considerable hydrological washing. In this case, the main factors of pedogenesis included humus accumulation, loessification, and claying. Upper levels of these paleosoils are disturbed by cryogenic processes and diagenesis, also characteristic of the commencement of the next Dnieperian epoch.

Warm interglacial epochs of the older half-period of the middle Pleistocene, which are correlative with oxygen-isotope excursions 11, 9, and 7, were different in intensity and types of pedogenesis. They were succeeded by a long period of considerable climatic cooling that resulted in one of the largest glaciations. This Moscovian or Dnieperian glaciation, as it is termed in different stratigraphic schemes, is well correlated with oxygen-isotope excursion 6 (Shik, 1995). Two older periods of cooling, which separate epochs of intense pedogenesis, correspond to excursions 10 and 8.

Outside the glaciation area in the middle courses of the Volga River, periglacial deposits of the maximum cooling epoch of the middle Pleistocene are 1.25–3.4 m thick, inhomogeneous in lithology, brown in color, and deprived of layering. The loess-like clayey loam of that period is studied in the Tatarskaya Chishma, Chirkovo, Tigany, Komintern, and other sections (Fig. 1). These notably calcareous ($CO_{2\,carb}$ = 3.9–4.26%) and humified (C_{org} = 0.32–0.37%) deposits have a lower content of the silty fraction (15.6–19.0%) than genetically similar sediments of older loess-accumulation periods (25.2–27.6%).

Late Pleistocene. Deposits of this age are widespread in the studied region of middle courses of the Volga River. In river valleys of the periglacial zone, the upper Pleistocene alluvial deposits are detected as constituents of the first, second, and third terraces above the flood plain (Shantser, 1935; Moskvitin, 1958, 1965; Goretskii, 1964). Subaerial deposits with paleosoil horizons in the watershed areas and high terraces are also of this age. In many sections, the respective loess–soil formation includes three horizons of fossil soils corresponding to the main phases of pedogenesis different in age and evolutionary trend (Glushankova, 1992, 1994). Their biostratigraphy and morphology, along with radiocarbon dating, suggest them to be successively corresponding to the following events: the Mikulinian (Eemian, Riss–Wurm, Sangamonian) interglaciation correlative with the lower part of oxygen-isotope excursion 5 (substage 5e); the first interstade of the Valdayan (late Volga, Amersfoort, Brörup) glaciation (Ivanova, 1966, 1980; Velichko, 1973); and the Bryansk (Danube) interstadial (Chebotareva and Makarycheva, 1974). Soil horizons of the loess–soil Valdai Group are separated by the Mezin, Kalininian, and Ostashkov horizons of loess-like clayey loam included by ISC into the regional stratigraphic scheme of 1984. The Kalininian cooling corresponds to oxygen-isotope excursions 5 (substages 5a–5d) and 4, whereas the Ostashkov Horizon is correlative with excursion 2. These events of climatic cooling are separated by a long epoch usually classified as megainterstadial, though the climate of that time was cooler than now. The oxygen-isotope excursion 3 corresponding to this event is less prominent (indicating not very significant warming) as compared to other odd excursions suggested to be indicators of interglacial epochs (Shik, 1995).

At the base of the loess–soil sequence of the late Pleistocene, which is located at the depth of 3.0–8.0 m below the surface, there is a polygenetic pedocomplex, whose stratigraphic position is the most persistent of the Pleistocene soils. In the reference sections, paleosoils of the Krutitsy (an upper portion of the pedocomplex) and Mikulinian interstadials are superimposed one onto another. This thick (1.5–3.05 m) pedocomplex of the composite structure is correlative with the Mezin pedocomplex of western and central areas of the loess province in the Russian plain (Velichko and Morozova, 1963). The warm epochs of pedogenesis were separated by a cooler period of loess accumulation. The loess bed 0.3–1.0 m thick is usually reworked by pedogenetic processes of the Krutitsy interstadial and seemingly represents the source stratum for the soil formation. It clearly divides the soil horizons only in the Chirkovo and Komintern sections.

The pedocomplex in question is confidently detected and correlated on the basis of diagnostic features even in sections far remote one from another and located in different geological and geomorphological areas of the Dnieper and Don river basins. The contrast features reflecting two phases of its formation history are quite distinct in the past forest zones and less clear in the past forest–steppe areas. In the middle Volga and

lower Kama river basins, processes of humus accumulation were more intense, and this resulted in a lesser differentiation degree of the same pedocomplex profile, especially in depressions of the past relief. In some sections, which are located in lower courses of the Kama River, the pedocomplex profile consists of horizons A_1, B_{Ca}, and C_{Ca}, the latter grading into the basal sediment. The soils in question experienced cryogenic deformations during the Smolensk phase of the early Valdai cryogenesis (Velichko, 1973).

The Mikulinian interglacial optimum corresponded to the epoch of intense pedogenesis, and the soil horizon that covered the studied region of the Russian plain presumably included several soil types, each of them characteristic of particular elements of the past landscape. The watershed areas of the plateau and hummocky types were covered with soils corresponding to several modern analogues: the gray differentiated eluvial–illivial soil of the Podzol type; the leached mycelium–carbonate Cherozem affected by podzolization; and the grassland Chernozem. Differentiated soil profiles with sequences of horizons A_1, A_2–B_t–C and A_2, B_t, BC, B_{Ca} were characteristic of small depressions in the past relief. Their horizon B_t is enriched in silty material and sesquioxides as compared to the basal rock. Most likely, these soils were formed due to a higher humidity in the depressions favorable for the pedogenesis of the eluvial–illuvial and eluvial–gleying types. They were also widespread, as at present, in the past forest–steppe landscapes. Fossil soils of the relief forms between depressions consist of horizons A_1, B, and C, suggesting them to be of the recent Chernozem type.

In some sections, the Chernozem-like soils of the Mikulinian time display a peculiar clayey horizon free of carbonates, which is situated between the humic-accumulative and calcareous horizons and was formed due to leaching. Distribution patterns of silt, clay, and sesquioxides in their profiles suggest a weak influence of eluvial–illuvial differentiation. This factor resulted in the appearance of humic leaks and brown films or crusts on the parting planes in the horizon B_t. Lower parts of these soil profiles may display gleying marks. The Chestnut soils, which may alternate in the sequence with the gray forest soil, have been insignificantly affected by podzolization, as is evident from the light silica substance dispersed in the structural parting fissures of the boundary zone between horizons A_1 and AB. This feature appears to be characteristic of soils formed in the past forest–steppe areas.

The Mikulinian soils similar to the recent mycelium–carbonate Chernozem of southern Europe are widespread in the lower courses of the Kama River. They are thick (up to 2 m), and their dark brown humic horizon (0.8–1.2 m thick and sometimes with a grayish tint) overlies the illuvial horizon of highly ferruginate clayey loam with carbonate pseudomycelium. Plentiful mole tunnels are typical of both units. According to analytical data, composite soil profiles of this type are rather homogeneous in chemical composition, although sesquioxides tend to be more concentrated in the humic horizon. The content of calcium-humate organic matter (C_{ha}/C_{fa} = 1.3–2.1) is quite variable (C_{org} = 0.14–0.53%), and its maxima are confined to the upper (0.53%) and lower (0.45%) humic horizons of two paleosoils constituting the pedocomplex. Distribution patterns of carbonates in the soil profile also reveal several zones of peak concentration: one is confined to the illuvial horizon of the upper soil, and two others are recorded in the humic and illuvial horizons of the lower soil. In the last case, the concentration maximum is confined to the upper part of the unit.

Many features described above are characteristic of Chernozem. However, humification in paleosoils of the Mezin pedocomplex is more intense than in the contemporary Chernozem horizon occurring in the lower courses of the Kama River. In the lower part of the profile, these paleosoils also contain small recrystallized concretions and pseudomycelium segregations of carbonates, indicators of the seasonal migration of the last compounds. Punctate Fe-Mn-films scattered in transitional horizons suggest that the soil-formation process progressed in humid environments. According to other characteristics, seasonal washing also affected the structure of soil profiles, although general environments of soil accumulation were of a more distinct continental type than in the present-day zones of the carbonate–mycelium Chernozem. For instance, this is evident from the fractured state of the lower soil boundary, which reflects recurrent desiccation and moistening. Some unstable characteristics of soils accumulated during the Mikulinian interglaciation probably reflect the microrelief influence.

In contrast, soils of the Krutitsy interstadial (upper members of the Mezin pedocomplex) are rather uniform. Their profiles include the dark gray to dark brown humic horizon (0.7–1.5 m thick), where organic matter is of a distinct humate composition. The humus content here (C_{org} = 0.32–0.53%) decreases downward, and this indicates the prime significance of sod formation in the history of the pedocomplex. The paleosoils in question are weakly differentiated with respect to their chemical composition. Morphological criteria, physicochemical parameters, and uniform features of organic matter characteristic of this soil-formation stage suggest a stable combination of elementary pedogenetic processes. The steady character of humus accumulation was responsible for the development of invariable soil profiles consisting of horizons A_1 and C. Consequently, we may assume that the soil horizon of the considered epoch of pedogenesis was homogeneous throughout the region of middle courses of the Volga River and thus reflected the hyperzonal environments of the respective interval of the late Pleistocene. Soil profiles of the Krutitsy interstadial are disturbed by the early Valdayan phase of cryogenesis (the Smolensk

phase "b" after Velichko, 1973); this is one of their diagnostic features.

Some researchers (Liberoch, 1964; Fink, 1969; Bronger, 1969) suggested these soils to be of the Chernozem type; however, the absence of carbonate illuvium in their profiles discredits such a viewpoint. The soils are highly humified and seem to be of the sod–chernozem genesis, which is untypical of recent soil horizons. Their origin in steppelike landscape areas is implied by the remains of piping hare, ground squirrel, and sagebrush vole recovered from mole tunnels (Markova, 1982).

Paleontological remains of the Mikulinian time from the Volga River region are known from the Barmino, Gorodets, and Elasy localities (Fig. 1). Microteriofauna of the first locality consists of voles of the *Clethrionomys, Arvicola, Microtus,* and *Lagurus* genera, which coexisted with hamsters, ground squirrels, insectivores, and piping, and real hares. Agadzhanyan obtained additional data on the Mikulinian fauna from the Elasy locality, where it includes *Clethrionomys glareolus* Schreb., *Cl. rufocanus* Sundey, *Microtus agrestis* L., *M.* (*Stenocranius*) *gregalis* Pallas, *Eolagurus*, and field mouse forms in addition to the aforementioned assemblage. As one can see, steppe species are rare here, whereas representatives of the forest fauna are preponderant and define the type of the community. Water vole from this assemblage resembles the recent species; however, this rodent had another structure of dental enamel identical to that of *Arvicola sapidus* Miller from the Iberian Peninsula (Agadzhanyan and Glushankova, 1989).

The interstadial soil-formation phase in the steppe zone was succeeded by the epoch of loess accumulation during the middle Valdayan cooling. Microteriofauna of this time is known from the Shakhmanovo locality, where it has been found in a member of cross-bedded alluvium of the second terrace above the floodplane of the Sundovik River (Fig. 1). The main group of this fauna consists of predominant sagebrush (*Lagurus*) and less abundant stenocranium and water voles. Other rodents are represented by lemmings and ground squirrels. The fauna composition indicates a degradation of forest communities and progradation of steppe biota in the Volga River basin during the second half-period of the late Pleistocene.

Inside the Valdayan loess sequence exposed in reference sections of the Volga and Kama river basins, there is a fossil soil occupying a higher level than that of the Mezin pedocomplex. According to the distinct stratigraphic position and diagnostic structural features, this soil horizon is referred to the Bryansk warm period. The upper age limit of the period is defined as equal to 22–24 ka on the basis of radiocarbon dating of humus specimens (Morozova, 1981).

Peculiar soils of the Bryansk stage differ from both the interstadial and the recent pedocomplexes. They are of the sod–gley–gelifluction genesis, and the structure of their profiles enriched in clay particles comprises either two (A_1 and B_{Ca}), or three horizons (A_1, B_{Ca}, and C_g). The variegated humic horizon varies in color from the yellow-brown upper part to the grayish light brown lower portion. It displays an intense gleying (local punctation and mottling) and some patches of organic matter penetrating far downward, even in the underlying horizon. Organic matter is dark brown in the uppermost interval of the horizon and medium brown below. The illuvial horizon is ferruginate and contains carbonate pseudomycelium and concretions, though the former is poorly developed.

These soils occurring in the Volga, Kama, Don, and Dnieper river basins are genetically interrelated according to the following criteria: (1) their humus is of the fulvate type (C_{ha}/C_{fa} = 0.20–0.51), the aromatic series of humic acids is combined with sesquioxides, the content of the nonhydrolized residue is high, and analytical data suggest a highly dispersed state of humic acid molecules indicating a considerable degree of carbonization and oxidation; (2) the illuvial carbonate horizon is located at the base of the profile, and indicators of gleying are not always distinct below it; (3) the soil profiles are enriched in the clay fraction.

Genesis of soils of the Bryansk time is not very clear and variably interpreted, because their analogues are missing in the present-day soil horizons. In addition, the Vladimir phase of cryogenesis at the end of the Bryansk period (Velichko, 1973) disturbed the soil profiles, stimulated the secondary gleying, and resulted in mixing of the soil mass. Some diagnostic criteria suggest that these soils were formed in continental environments under the influence of the terminal cold phase and local geographic position responsible for the provincial gradation of moistening. The pedocomplex under consideration is almost hyperzonal in the structure, and respective humus accumulation seems to have been combined with internal weathering, gleying, and carbonatization. These soil-formation processes are now characteristic of extracontinental climatic zones, such as cryoarid regions of Siberia, for instance, central Yakutia (Sokolov *et al.*, 1979; Sokolov, 1980).

The middle Valdayan interstadial was relatively warm and preceded the cold and arid late Valdayan epoch of accumulation of periglacial deposits. In the Oka and Sura river basins, these deposits are 2.6–3.5 m thick and composed of calcareous ($CO_{2\,carb}$ = 1.97–2.39%) loess-like loam, whereas in the lower courses of the Kama River they are represented by calcareous ($CO_{2\,carb}$ = 2.3–4.2%) alluvial beds, which are widespread, 1.2–2.6 m thick, and light brown in color. The deposits are inhomogeneous in their granulometric composition: medium- to fine-grained sand and silt are predominant in the topmost and basal portions of their profile, whereas its middle part consists of clayey silt. Distribution of major chemical components reflects the lithological heterogeneity: the maximum content of sesquioxides is confined to the middle interval of the

profile, whereas silica is concentrated higher and below this interval. Organic matter of fulvate composition is not very abundant (C_{org} = 0.13–0.29%) and homogeneously dispersed in the soil (Glushankova, 1992).

The post-Valdayan (early Holocene) stage in paleogeographic evolution of the middle Volga River basin resulted in accumulation of gray-brown ferruginate loam and sand of the oxbow-lacustrine facies with horizontal bedding. These deposits are 2.1–3.1 m thick in the studied Tigany and Attikovo sections (Fig. 1). According to granulometric parameters, aleuritic (22.4–37.1%) and fine- to medium-grained sandy fractions (18.6–32.4%) are predominant in the rock composition. The beds are rather rich in organic matter (C_{org} = 1.0–4.2%) but depleted in carbonates ($CO_{2\ carb}$ = 0.56–1.61%). Distribution of other chemical components is relatively stable. Calcium sulfates are slightly concentrated at the top of the sequence and represent principal water-soluble compounds. Agadzhanyan, who studied microteriofauna from these deposits, considered it as corresponding to the typical interglacial assemblage. Along with dominant common field and forest voles, it includes field mice, abundant screws, and small Mustelidae. Water vole is not abundant like the characteristic steppe taxa (ground squirrels, sagebrush voles, and northern moles). *Microtus arvalis* Pallas and *M. agrestis* L., neither being frequent in older communities of the Volga middle reaches, are rather common at this stratigraphic level. Many species of the early Holocene time do not differ in morphology from their modern analogues, but the general composition of the assemblage is distinct. For instance, out of twenty-seven species of the Tigany fauna, seven do not exist in the region at present. Their distribution areas are now located southward and to the east of the middle courses of the Volga River. Using these data, Agadzhanyan concluded that small mammals of the Tigany and Attikovo localities dwelled in a warm climatic zone resembling that of the present time. However, the water system in the region was more vivid, and biotopes were more diverse and contrasting. Summer temperatures presumably were higher.

CONCLUSION

Thus, a comprehensive study of reference sections in the glacial and periglacial provinces of the middle Volga and lower Kama river basins revealed their principal lithological, geochemical, paleopedological, and paleontological characteristics. The data obtained are used for distinguishing facial–genetic stratigraphic horizons and for their chronological correlation with the Pleistocene stratigraphic scale on the basis of microteriofauna and pedological criteria. These data are also used to depict trends in paleogeographic evolution of the region during the Pleistocene and Holocene. They provide a deeper insight into the stratigraphy and paleogeography of the East European part of the Russian plain, where Recent deposits, especially the loess–soil formations, had been inadequately studied before.

(1) The research data show that the Pleistocene history of the studied region comprises several important bioclimatic events: the pre-Okaian and Okaian–Dnieperian warming stage including the Likhvinian, Kamenka, and Romny interstadials; the Dnieperian glaciation; the Mikulinian interglaciation; the middle Valdayan periglacial stage; the late Valdayan cooling; and early Holocene warming. The Likhvinian and Mikulinian interglaciations consisted of climatic rhythms of the second rank.

(2) Directed and peerless changes in climatic environments at different stages of the Pleistocene paleogeographic evolution resulted in accumulation of soil horizons rhythmically alternating with beds composed of the loess-like loam or corresponding to alluvial facies. The climatic fluctuations are also recorded in pedogenetic, lithological, and geochemical peculiarities of soil horizons and other recent deposits of the region.

(3) The Pleistocene history of the middle Volga and lower Kama river basins comprises at least eight epochs of intense pedogenesis unequal in duration and diverse in climatic conditions. Each of them was unique and produced stratigraphic horizons with different soil types variably distributed in the region in response to evolutionary changes in the Pleistocene landscapes. There were six cold stages of pedogenesis, and alluvial deposits corresponding to the two oldest stages yielded microteriofauna with boreal species.

(4) Soil horizons of the interglacial Pleistocene intervals include a complex zonal spectrum of soils formed in the present-day mode over the entire region of middle courses of the Volga River. The paleosoils of watershed areas were accumulated under conditions of forest, forest–steppe, and steppe landscapes.

(5) The results obtained show that, in the older half-period of the early Pleistocene, the pedocomplex of the interglacial optimum mainly consisted of Chernozem-like, grassland Chestnut, and dark grassland soils. Soils of the Likhvinian interglacial optimum were accumulated in subboreal forest and steppe landscapes, and, in the last case, they also went through the grassland stage of evolution. Pedogenesis of this period included several elementary processes, such as loessification, podzolization, claying, and humus accumulation. Brunizem and Chernozem-like soils were characteristic of the post-Likhvinian stage, when humus accumulation represented the most important process of pedogenesis. The next, middle Pleistocene, epoch resulted in accumulation of soils typical of semiarid steppe areas of subtropical and temperate climatic zones. Relatively arid continental environments of soil formation appeared in the late Pleistocene. Interstadial soils of this period are the products of sod formation and gleying–gelifluction processes. Pedogenesis in the forest and steppe landscapes of the Mikulinian interglaciation

is of the type characteristic of the temperate climatic zone now.

(6) The Pliocene–Pleistocene microteriofauna from the middle courses of the Volga River represents a transitional zoogeographic link between European and Asiatic faunas; this link is useful for the biostratigraphic correlation of respective deposits in the western and eastern Paleoarctic provinces. Asiatic taxa existed among small mammals of the studied region until the Holocene, although their significance gradually decreased. Boreal species appeared here as soon as in the Pliocene. Afterward, they occurred in the first half-period of the early Pleistocene and then in the middle–late Pleistocene and early Holocene as indicators of cold epochs in the paleogeographic evolution of the region.

(7) The Pleistocene trends of paleogeographic evolution of the middle Volga–lower Kama and Dnieper–Don river basins are similar in their basic features.

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Boris Maksimovich Keller (1912–1997)

Boris Maksimovich Keller, doctor of geological–mineralogical sciences, professor, a prominent figure in geology with broad scientific interests, who devoted a good part of his life to stratigraphy, died on January 14, 1997. On October 14, 1997, he would have been 85 years old.

Keller began his geological research in 1933, when he was a student at the Moscow Geological Prospecting Institute. After graduation, he worked in Moscow, first, at the Oil Research Institute (1933–1936) and then, for 50 years (up to 1985), at the Geological Institute of the USSR Academy of Sciences (GIN), where he started his research activity under the guidance of V.V. Menner. In 1956, with the active support of N.S. Schatsky, he organized and headed a research team to investigate the Upper Precambrain stratigraphy. Later, together with his team, he founded the Laboratory of Upper Precambrian Stratigraphy.

Keller was known as a versatile and always successful researcher. In the 1930s, he was involved in investigating foraminifers and the Cretaceous and Paleogene stratigraphy in oil provinces of the southern Soviet Union. In the 1940s, he turned his attention to coal deposits and rock formations of the Urals. In the 1950s, he investigated the Ordovician stratigraphy and graptolites in Kazakhstan, and, from the 1950s onwards, until his retirement, he studied the stratigraphic formations, organic remains, and geological history of the Late Precambrian. During his 50-year-long professional life, he also dealt with general problems of geology such as geological formations, theoretical problems of stratigraphy, paleogeography, and scientific information. When he was 27 years old, he defended his candidate thesis in geology, and, at the age of 36, he earned his doctorate. He has authored about 150 publications, among them several monographs. Thanks to his rich experience, erudition, and great energy, Keller was recognized as one of the leading geologists and stratigraphers in the Soviet Union. For his significant contributions to science, he was awarded the Order of the Red Banner.

Keller was always interested in the most vital problems ("inciting," as he called them), the solution of which was necessary and urgent for further progress in geology and stratigraphy, no matter to which geological era they were related. He tried to apply a combination of techniques to stratigraphic research, and, having an exceptional ability to organize, he managed to gather various supporters: biostratigraphers, experts in different fauna and flora groups, geochronologists, lithologists, regional geologists, etc. These scientific groups usually united specialists from GIN and other institutions, both central and regional. The vast factual material, meticulously collected and interpreted by those informal research groups, was useful for obtaining not only new important results on the regional stratigraphy and geology, but also general scientific overviews. The investigations headed by Keller usually culminated in fundamental publications such as monographs or, at least, series of scientific articles, which became important in solving particular problems.

This style of research is well illustrated by the elaboration of a complex approach to the Upper Precambrian stratigraphy and creation of a standard stratigraphic scale for the Upper Precambrian of the Soviet Union under Keller's leadership. The work was carried out at the GIN Laboratory of the Upper Precambrian Stratigraphy with the assistance of specialists from other institutions. The objective of the work included the following: first, the study on stromatolites, microphytolites, and microfossils; second, the isotopic age determination of deposits; and, third, the investigation

of paleogeography and tillites in nearly all major sections of the Soviet territory, including the Urals, Russian and Siberian platforms, and Central Asia. Later, the study involved a number of important sections in foreign territories: in Spitsbergen, the United States, Canada, Australia, People's Republic of China, Poland, India, and England. The investigations finally resulted in a series of monographs by Keller and other participants on the regional stratigraphy and methods of stratigraphic research of the Upper Precambrian, including the fundamental book entitled *The Upper Precambrian Stratigraphy* (one in the multivolume series *Stratigraphy of the USSR*) written under his leadership, and the first standard stratigraphic scale of the Upper Precambrian in the USSR (1979), which served as a basis for the geological mapping of areas with Riphean and Vendian deposits in this country. These distinguished works were really an epoch in Upper Precambrian research.

Studies on the Upper Precambrian that began as far back as 1952 turned into Keller's life-long interest. He proposed the three-member subdivision of the Riphean, which is now generally recognized. He made the first important finds of Vendian–Ediacarian fauna in the Russian platform, thus opening wide opportunities for studying this unique period of paleontological records. He enhanced the elaboration of the historical–geological method of subdividing the Precambrian, substantiated the historical–geological boundaries of general subdivisions of the Riphean, and clarified their specific features. Works by Keller served as a methodological basis for the recent chronostratigraphy of the Precambrian. It is essential that, on the one hand, he demonstrated how to apply a combination of geological, paleontological, and physical methods for solving stratigraphic problems and, on the other hand, how to evaluate the geological significance of isotopic–geochronological, paleoclimatic, or tectonic evidence in stratigraphic relations observed. Keller directed and participated in compiling paleogeographic and paleotectonic maps of the Riphean and Vendian of the Soviet Union, unique at that time, and contributed to studying the structure, paleontology, and isotopic age of the Riphean stratotype.

Keller successfully combined his scientific research with teaching activities; he encouraged his postgraduate students and those from other institutions to participate in his research. These investigations served as a basis for their candidate and doctoral theses. Keller carefully followed the process of their work and discussed the progress and achieved results either individually or at the laboratory meetings, criticizing, inspiring, and speeding up the authors. Everybody received good advice, moral support, and stimulating words. During his long professional life, Keller prepared several dozen candidates and doctors of science. Involving young specialists in research activities, he helped them to become real experts in stratigraphy and geology. He truly deserved his rank of professor. In 1968, Moscow State University published his book *Upper Proterozoic of the Russian Platform (Riphean and Vendian),* which was in fact an educating course.

Keller also had an exceptional organizing skill. He was one of most active members of the Interdepartmental Stratigraphic Committee (ISC), and, for 30 years, he served as the chairman of the ISC Permanent Commission on the Upper Precambrian. During that period, he initiated and participated in several all-union conferences on the Precambrian that resulted in the successful release first of the regional and then general stratigraphic scales of the Precambrian deposits in the Soviet Union. He was also involved in the All-Union Certifying Commission, Library Committee, and other numerous scientific committees and commissions of the USSR Academy of Sciences.

Keller was associated with the All-Union Institute of Scientific and Technical Information (VINITI) from its founding up to the 1980s, serving as an editor of the sectors *General Geology* and *Stratigraphy, Paleontology* of the periodic abstract volume *Geology.* He participated in developing the concept and structure of the information on earth sciences and conducted the editorial work. The high scientific status of the periodical, very popular in the geological community, was maintained mostly due to Keller's activity.

Keller was also active in other editorial activities. He published a number of popular scientific brochures and articles, wrote and reviewed geological essays for encyclopedias, and edited translations and participated in translating some fundamental works in geology into Russian, among them monographs *Principles of Stratigraphy* (1962) by K. Danbar and J. Rodgers, *Stratigraphic Geology* (1952) by M. Gignoux, and selected papers by Chinese geologists. He edited and assisted in publishing several dozens of books in the *Nauka* and *Nedra* publishing houses, in VINITI, and other institutions. Most of them were devoted to vital problems of stratigraphy.

Many works by Keller are still in active use, though they were written many years ago. It is impossible to refer to them all, but we shall mention some of them. Nowadays, micropaleontologists benefit from Keller's works on the Cretaceous published as far back as the 1930s. These works contain all elements of the modern biostratigraphic analysis: the study on planktonic and benthic foraminifers; an excellent exposition of the regional stratigraphy; and subglobal correlations with due regard to biogeographic, climatic, and facies peculiarities. In addition, these works were among the first to manifest the important role of micropaleontology for oil geology. The last issue fostered an advance of this biostratigraphic branch in our country. Among his other works widely used at present, there are those written in the 1940–1970 period on the Riphean flysch series and the interpretation of geodynamic settings during their accumulation, as well as papers on the evolution of geological formations in the Earth's history. Keller

suceeded in identifying the allochthonous terrigenous formations, although this achievement was not duly appreciated until recent time.

The main scientific heritage of Keller is certainly his contribution to the Upper Precambrian stratigraphy and his scientific school in this branch, which still occupies a distinguished place among national geological schools.

Keller was very enthusiastic. Far from being an armchair scientist, he was an experienced field geologist. With his geological hammer and rucksack, he covered practically the entire Soviet Union. Among the regions he preferred to work in are the Urals, Crimea, and Kazakhstan. His last field season was in the Urals, when Keller was 72 years old. Two years before that, he participated in field works in the Kharaulakh Mountains of northern Yakutia. Keller was a kind-hearted and optimistic person with great and diverse knowledge. Until his last days, he was interested in belles-lettres, art, and history. To talk with him was pleasant, interesting, and beneficial.

He will be remembered not only for his great contribution to science. He will remain in the hearts of those who knew him as an inspiring teacher and colleague.

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Grigorii Yakovlevich Krymgol'ts (1907–1997)

Grigorii Yakovlevich Krymgol'ts, professor of the Chair of Historical Geology at St. Petersburg University, passed away on October 2, 1997.

With his death, science has lost an outstanding stratigrapher and paleontologist and a distinguished expert in the Jurassic System, particularly in ammonites and belemnites.

Krymgol'ts was born on January 20, 1907, in the city of Simferopol. In 1924, he moved to Leningrad and entered the division of soil science of the Physical–Mathematical Faculty at Leningrad State University (LSU) and completed his education in 1929 together with the other first graduates of the Soviet Union (about 30 students), young specialists in geology. Many of them were stratigraphers and paleontologists (O.S. Vyalov, A.G. Eberzin, N.P. Lupov, V.S. Slodkevich, and others). As a student, Krymgol'ts began to work at Geolkom, afterwards the All-Union Institute of Geology (VSEGEI), and was associated with this institute until the end of his life. In 1932, he began to teach at the Geological Faculty of LSU, and, in 1948, he became an associate professor at the LSU Chair of Historical Geology. In 1965, after successfully gaining his doctorate in VSEGEI, he received the position of professor. He left the university in 1994 because of his poor health, though he never broke connections with the Chair of Historical Geology until the end of his life.

An expert in stratigraphy and experienced field researcher, Krymgol'ts was very interested in general problems of stratigraphy, particularly, in the category of zonal subdivisions commonly used in global and regional stratigraphy. A quarter of a century ago, he introduced the new concept of "Lona" for a zone of paleobiogeographic provinces; this idea became widely used in Phanerozoic stratigraphy and was reflected in the system of stratigraphic subdivisions of the Stratigraphic Code of Russia.

Teaching at the university beginning in 1934, Krymgol'ts educated and trained numerous students. Many of them have become prominent geologists, professors, and members of various academies. He was an outstanding teacher with an exceptional gift for bringing his great knowledge to others. He went deep into the research work of his students and colleagues and assisted them in writing diploma papers, candidate and doctoral theses, articles, and books.

Krymgol'ts guided the research of 24 postgraduate students from Russia, Ukraine, Georgia, Azerbaijan, Armenia, Turkmenistan, Uzbekistan, Tajikistan, Romania, Bulgaria, Germany, and Vietnam who successfully defended their dissertations of candidate of science.

His high scientific authority was recognized worldwide. He was the chairman of the Permanent Commission on Jurassic Stratigraphy of the Interdepartmental Stratigraphic Committee of the Soviet Union (Russia) and a member of its bureau. Krymgol'ts was also a member of the International Commission on Jurassic Stratigraphy, a honorary member of the Council of the All-Union Paleontological Society, a member of the Council on Paleontological Problems at the USSR Academy of Sciences, a member of the Presidium and Section on Geology and Mineralogy at the Leningrad (St. Petersburg) Society of Naturalists, and a member of the scientific boards of the Geological Faculty at Leningrad University and VSEGEI, where he also worked as an expert of special doctoral boards. He was the organizer and scientific leader of the Turkmeni Expedition of LSU, whose activity was in progress from 1956 to 1991.

Krymgol'ts is the author of more than 150 works. His major articles and monographs are as follows:

Upper Jurassic Cylindroteuthinae of the Timan and Sysola River Basins and the Orenburg Province (1929), *Jurassic Belemnites from the Crimea and Caucasus* (1931), *Materials on the Jurassic and Lower Cretaceous Stratigraphy of Turkmenistan* (1934), *Lower Cretaceous Belemnites of the Caucasus* (1939), *Materials on Marine Jurassic Stratigraphy of the Bureya River* (1939), *Stratigraphy and Faunas of Mesozoic Marine Deposits in Northern Siberia* (1953), *Methods of Identifying Mesozoic Cephalopods* (1960), and *Ammonites from the Lower and Middle Jurassic Deposits in the Northern Caucasus* (1961). Moreover, he is an author and editor of several other well-known books: *Geology of the USSR* (vol. IX, *Northern Caucasus*, editions of 1947 and 1968; vol. XXII, *Turkmenistan*, editions of 1957 and 1972); *Geological Structure of the USSR*, vol. I, 1968; *Atlas of Guide Fossil Fauna of the USSR*, vols. VIII, 1947, and IX, 1949; *Fundamentals of Paleontology: Cephalopod Mollusks* (vol. II), and *Ammonoidea and Coleoidea*, 1958; *Atlas of Paleogeographic Maps of the USSR*, vol. III: *The Mesozoic*, 1961; *Stratigraphy of the USSR: The Jurassic System*, 1972; *Jurassic Zones of the USSR*, 1982, translated into English under the title *The Jurassic Ammonite Zones of the Soviet Union* by the Geological Society of America in 1988.

The great contribution of Krymgol'ts to science was rewarded by a number of governmental orders and medals.

Krymgol'ts was a good-humored, well-liked person; it is no surprise that he was always surrounded by people of younger and older generations. Everybody received his kind advice, necessary consultations, explanations, and encouraging words. Apart from stratigraphy, he taught his students and colleagues to keep to ethic principles in science and life. The best example here is his own long and fruitful professional life, devoid of vanity and self-interest. He freely gave his knowledge and expertise to his followers and colleagues.

With the death of Krymgol'ts we have lost a prominent geologist, outstanding stratigrapher (the chief "lawyer" in this science), excellent teacher, and unfailing friend. He is no longer with us, but we shall remember him as a man of unique intelligence, strong principles, human kindness, and gracious good will. He will be remembered by thousands of graduates of St. Petersburg University, paleontologists and stratigraphers who worked in Russia and foreign countries and who had contact with him and benefited by his expertise or advice.

B.S. Sokolov, A.I. Zhamoida, V.A. Prozorovskii,
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